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# REVIEW OF HYDROGEOLOGY RESEARCH TECHNIQUES AND TECHNOLOGY

by

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# REVIEW OF HYDROGEOLOGY RESEARCH TECHNIQUES AND TECHNOLOGY


## ABSTRACT

A review of techniques and technology pertaining to the movement of ground water, solutes/radionuclides and heat through porous and fractured media has been conducted. The theory describing each of these processes has been presented in terms of their partial differential equations. The equations serve as the basis for the identification of processes. These parameters have been discussed in detail with regards to their importance in controlling the overall transport processes. A hypothetical research program has been assembled for the purpose of illustrating the hydrogeologic methods and research techniques applicable to site characterization studies. Areas where the current state of the art is lacking have been identified and the necessary research has been recommended.

## RÉSUMÉ

On a effectué un examen des techniques et de la technologie relatives au déplacement des eaux souterraines, des radionucléides en solution et de la chaleur à travers des milieux fissurés et poreux. La théorie qui décrit chacun de ces processus a été exposée en ce qui concerne leurs équations différentielles partielles. Les équations servent de base à l'établissement des paramètres nécessaires pour quantifier le processus d'écoulement et de transport. L'importance de ces paramètres pour contrôler l'ensemble de ces processus de transport est traitée en détail. Un programme hypothétique de recherche a été élaboré afin d'illustrer les méthodes hydrogéologiques et les techniques de recherche applicables aux études des caractéristiques des sites. Les domaines où n'est pas suffisante la connaissance actuelle ont été identifiés et la recherche nécessaire a été recommandée.



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# 1.0 INTRODUCTION

The purpose of this report is to provide a review and evaluation of current hydrogeology research methods. The review will be structured in much the same manner as a research program proposal; that is, from underlying concepts and principles through to "proposed" methods of investigation. A discussion of the underlying hydrogeologic concepts and processes is provided in Section 2. The processes of ground water movement, solute or radionuclide transport and heat transport are discussed by employing the governing equations and the related field and laboratory research techniques and parameter evaluation methods. Interpretive techniques are addressed in Section 3 in the context of regional characterization and site characterization studies related to radioactive waste management or disposal site assessment requirements.

Numerous research methods are currently in use in the field of hydrogeology. This report will focus on, but not necessarily be restricted to, those methods the authors believe have potential for providing data or interpretations relevant to an assessment of a physical hydrogeologic flow system. However, site characterization, including ground water flow regime definition, must be accomplished using a number of data sets, of which the physical hydrogeologic parameters and measurements constitute only one. Other data sets such as hydrogeochemical analyses and the geologic, mineralogic and structural framework must clearly all be utilized to develop the most reasonable interpretation. This report addresses geologic systems from the points of view of porous and fractured media. The main differences in conceptualization and mathematical treatment are addressed where applicable. If research methods are applicable for one

and not the other, these differences are explained where necessary. If there is no distinction drawn, it can be assumed that the research method is applicable to both porous and fractured media although interpretive techniques may differ somewhat.

Hydrogeology has developed as an engineering and geology subdiscipline, starting with the empirical laboratory studies on sand columns by Darcy (1856). The results of these experiments have been generalized as Darcy's law, which relates the flow rate, Q, through a porous medium to the hydraulic gradient, $\Delta h/\Delta L$, the cross-sectional area of flow, A, and an empirical constant of proportionality, K, called the hydraulic conductivity.

$$Q = AK \frac{\Delta h}{\Delta L} \qquad (1.1)$$

Darcy's law remains the cornerstone of hydrogeology. The governing equations for ground water flow, discussed in Section 2, employ Darcy's law to obtain a parametric mathematical description of flow and solute or heat transport. Darcy's law was developed in the context of porous media; however, it has been applied almost universally to fractured rock situations, particularly with respect to water supply investigations, as they are primarily concerned with relatively large volumetric flow rates. The study of fractured media, as distinct from porous media, essentially began as a hydrogeology subdiscipline around 1960 (Barenblatt et al, 1960). Since that time, it has become evident that there are many instances in which the assumptions inherent in porous media developments are not applicable to fractured rock situations. However, there are also many instances where, with adequate parameter definition and scale considerations, porous media concepts can be adequately applied to fractured rock situations.

A schematic qualitative representation of the research and development of the hydrogeologic "knowledge/data base" is shown in Figure 1.1 starting with Darcy's empirical experiments in 1856. A significant change in slope of the curve occurs with the development of a quantitative description of ground water flow based on an analogy with heat flow (Theis, 1935). With this development the understanding of the ground water movement processes and nature of aquifer systems greatly increased. A second major development occurred shortly thereafter with Hubbert's (1940) conceptual model of a ground water flow system and his quantitative description of the energy distribution within a ground water system. Hubbert determined that the fluid potential, or the mechanical energy per unit mass of fluid, is directly related to the water levels in wells. The water level, or hydraulic head, in a well represents the sum of the elevation and the hydraulic pressure corresponding to the point of measurement (i.e., the screened interval). Hubbert demonstrated a simplified case shown schematically on Figure 1.1 (Flow System Conceptualization) in which ground water recharge occurs in the vicinity of topographic highs and ground water discharge occurs in the adjacent topographic lows.

A further advancement occurred with the development and application of boundary - value mathematical techniques (Toth, 1962, 1963) to flow system analysis, while the last 15 to 20 years has seen computer-implemented numerical solution methods remove many of the previous limitations on quantitative analysis (e.g., Freeze and Witherspoon, 1966, 1967, 1968; Lafleur and Lantz, 1984). These developments have led to the conceptualization of ground water systems illustrated schematically in Figure 1.2.

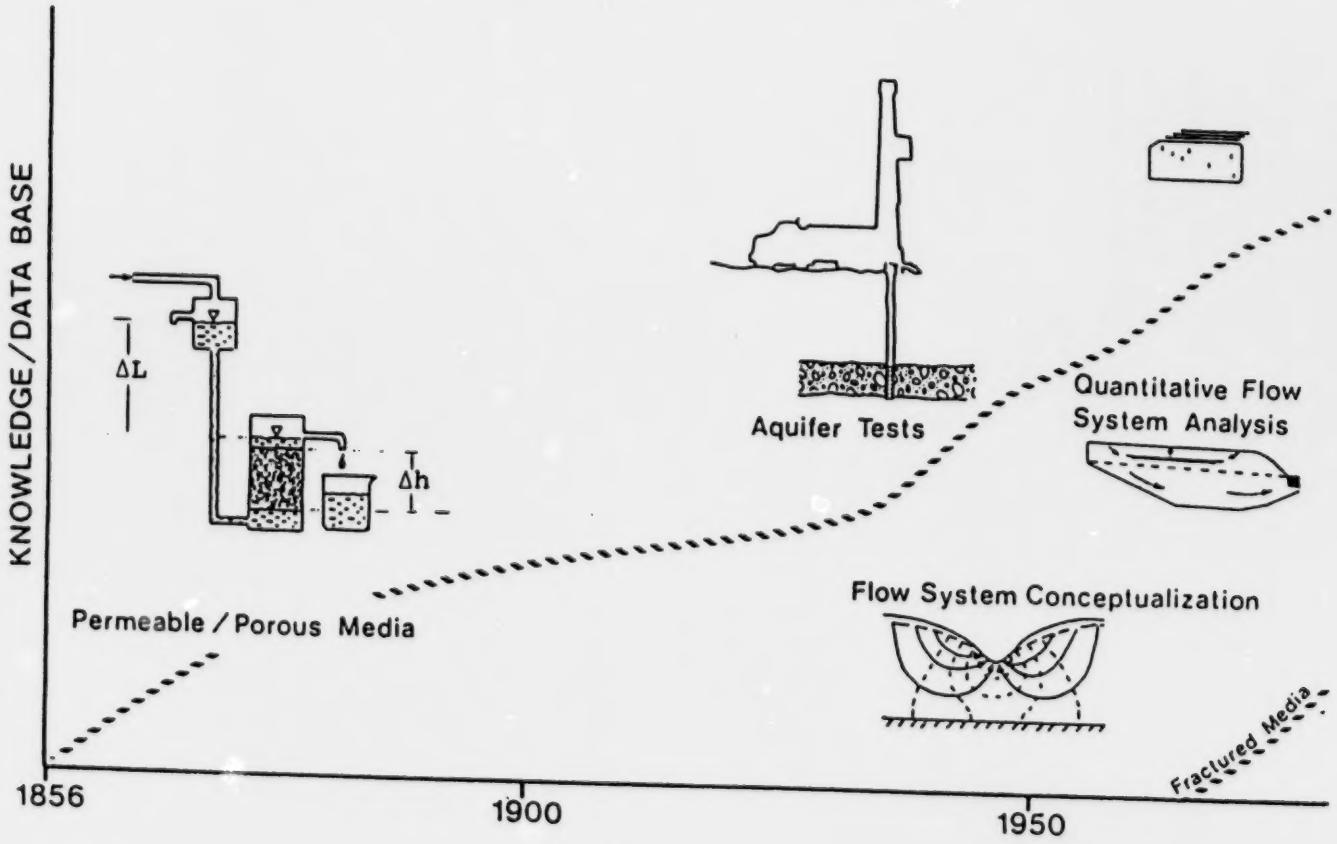
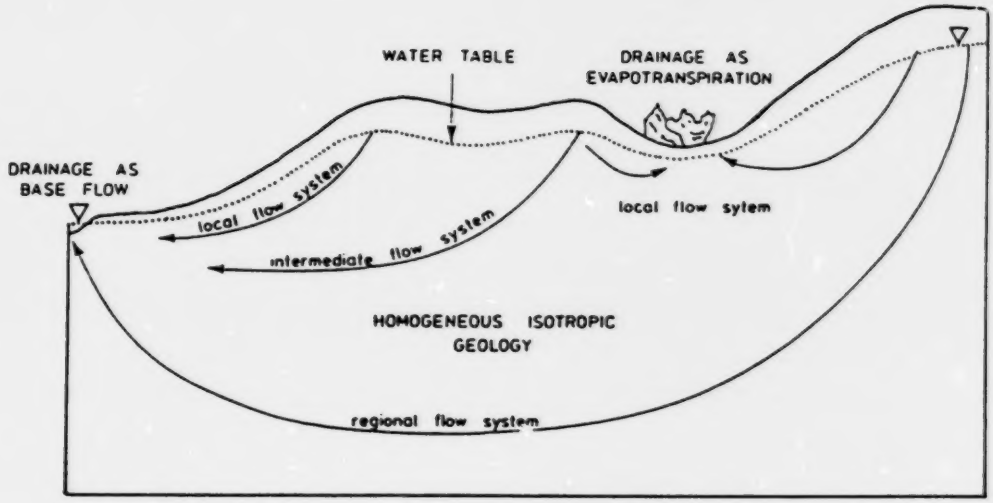
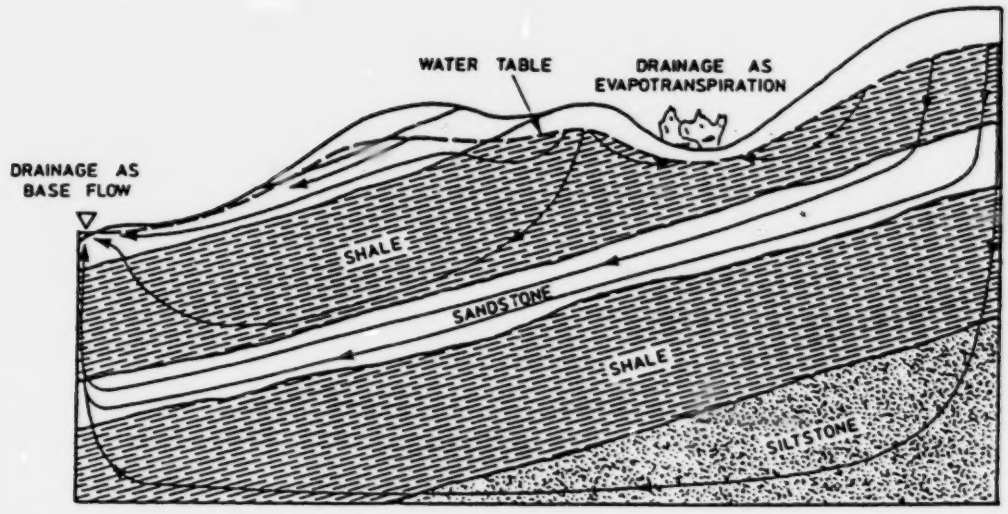



Figure 1.1 Historical Review of Hydrogeologic Methods Development




a)




b)

Figure 1.2 Ground Water Flow Systems a) in a homogeneous, isotropic geologic system b) in a subsurface system with variable geology and hydraulic conductivity

Ground water systems are usually conceptualized on the basis of scale. Toth (1963) defined three scales of flow system: "local" ground water flow systems are those in which the topgraphic highs that form the recharge areas are adjacent to the topographic lows that form the discharge areas; "intermediate" flow systems have recharge areas that are not necessarily adjacent to one another; "regional" flow systems have their recharge areas in the topographically highest point in the basin and their discharge areas in the topographically lowest points (e.g., the stream or river draining the basin).

The flow system simplification shown in Figure 1.2a assumes a homogeneous and isotropic (without directional properties) geologic system. Freeze and Witherspoon (1967) have quantitatively illustrated the effect on the flow patterns of variable subsurface geology, which results in variable hydraulic conductivity distributions. The same size system as Figure 1.2a is shown in Figure 1.2b with variable geology. The significance of geology and hydraulic conductivity variations is clearly evident.

The conceptualization of ground water flow, in terms of different scales of flow systems and variable geology, (Figure 1.2) assumes that the physics of ground water movement can be described using the principle of conservation of mass; or a material balance within a given three-dimensional volume. The basic aim of most hydrogeologic investigations is to establish the flow system at the scale of interest. This usually includes determination of the geologic framework within which ground water flows as well as the hydrogeologic characteristics of the geologic system.

Continuum analysis is considered to be the only feasible approach to evaluation of the consequences of radioactive waste emplacement; even though water flow, radionuclide and heat transport are controlled in some instances by fracture flow mechanisms. Studies of single fractures or fracture networks will continue to provide valuable insight into the processes which govern flow and transport in fractured media (An assessment of fracture zones and fracture permeability is provided in Appendix 1); however, the governing equations which are utilized for porous media analysis also apply to fractured media, provided the scale of evaluation is taken into account when the parameters are assigned definitions. This further requires that the parameter measurement is conducted at the appropriate scale, or that relationships exist which account for any scale dependency of the parameters.

## 2.0 HYDROGEOLOGIC CONCEPTS

The conceptualization and understanding of ground water flow systems requires a knowledge of the processes and assumptions that are inherent in the conceptualization. Most hydrogeologic research is conducted for the purpose of evaluating the concepts, verifying the processes or applying the measurement and interpretation techniques to specific characterization studies. This section will discuss the relevant concepts and processes related to ground water movement, solute or radionuclide transport, and heat transport and address the research techniques that relate to these areas.

The reader will rightly perceive a bias toward deterministic methods as opposed to probabilistic or stochastic methods for characterizing ground water systems. However, probabilistic or stochastic methods have recently become a topic of research interest in hydrogeology, particularly in the context of radionuclide transport. Most of the stochastic work to date has been directed at defining and employing a probability density function for the classical hydrogeologic parameters. Deterministic models are used to assess uncertainty by employing a statistical scheme to sample input variable distributions and performing repeated calculations to deduce statistical properties of the output variable range. Recent studies (e.g., Gelhar and Axness, 1983; Winter et al., 1983) have attempted to treat flow and solute transport in hydrogeologic systems stochastically, which requires restructuring the deterministic governing equations into stochastic equations. Although considerable progress is being made, a comprehensive stochastic description of hydrogeologic systems has not yet been developed. A detailed review of current methods and models, including deterministic, probabilistic and stochastic approaches, is in preparation under separate contract for the Atomic Energy Control Board.

## 2.1 Ground Water Movement

Ground water movement or flow occurs in response to a hydraulic potential gradient (Figure 2.1). Potential is defined (Hubbert, 1940) as " a physical quantity, capable of measurement at every point in a flow system, whose properties are such that flow always occurs from regions in which the quantity has higher values to those in which it has lower values, regardless of the direction in space". Hubbert showed that a relationship existed between the hydraulic potential and the water levels measured in wells in a ground water system. The water level in a well, or the hydraulic potential, is the sum of the elevation of the well screen and any hydraulic pressure which exists at the well screen which would cause ground water to rise to a level in the casing above the screen elevation. Essentially, according to Hubbert's definition of potential, ground water can move in any direction in the subsurface, with that direction determined primarily by the hydraulic potential gradient. The effect of anisotropy, or the directional properties of the medium, will be discussed later.

The amount of ground water that flows from point to point is determined by the hydraulic conductivity of the geologic media. The terminology related to the movement of ground water is somewhat confusing, with terms like hydraulic conductivity, permeability, intrinsic or specific permeability, coefficient of permeability, etc. In recent years, it has become common to simply refer to the term permeability as the "relative ease of water movement through a porous or fractured medium" and to employ terms such as "high permeability" to indicate the capability of high volumetric flow rates and "low permeability" to indicate a more limited capability. Permeability in a geologic medium results from the existence of interconnected, fluid-filled pore spaces through which ground water can move (Figure 2.2). These pore spaces can exist as holes between clay, silt, sand, or gravel sized particles, referred to as

## Ground Water Flow

- Occurs in Response to a Hydraulic Potential ($\Phi$) Gradient

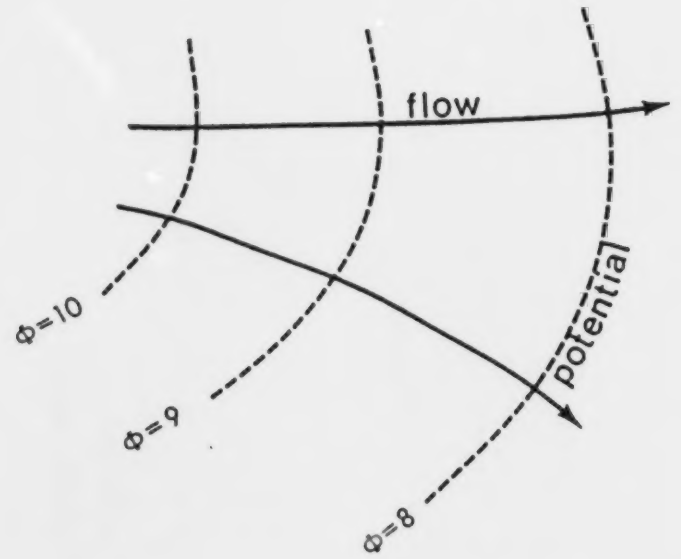
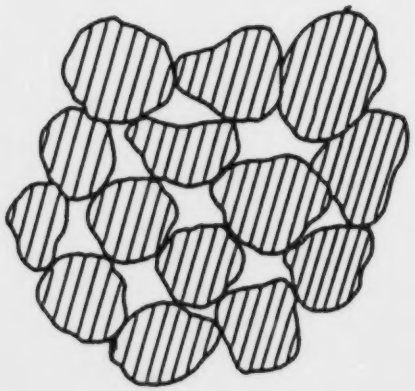



Figure 2.1 Ground Water Flow Occurs in Response to a Hydraulic Potential Gradient

# *Permeability*

**Relative Ease of Water Movement**

- Requires Interconnected, Fluid-filled Pore Space



Intergranular Pores

Fracture Pores

Figure 2.2 Illustration of the Nature of the Pore Spaces Contributing to Permeability in Porous and Fractured Media

"intergranular" pores, or as interconnected fractures in a more massive rock structure. The basic distinction between porous and fractured media is therefore the type of pore space which provides an interconnected porosity for water movement.

For clarification, the following definitions (Lohman et al, 1972) relate to the relative ease of water movement through geologic media.

hydraulic conductivity ($LT^{-1}$) -

> a function of the medium and the fluid, the hydraulic conductivity is the volume of water at the existing kinematic viscosity, that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow.

specific or intrinsic permeability ($L^2$) -

> a function of the medium only, the intrinsic permeability is the volume of liquid of unit kinematic viscosity that will move in unit time under a unit potential gradient through a unit area, measured at right angles to the direction of flow.

The term permeability (without the associated words "specific" or "intrinsic") is used in this report to refer to the concept of relative ease of water movement through a medium, rather than as the parameter defined above. The term hydraulic conductivity is used as both a parameter with units $LT^{-1}$ and as a concept identical to that of permeability.

Ground water flows directly down gradient, or at right angles to the contours of hydraulic potential in a completely isotropic system (see Figure 2.1). However, most hydrogeologic systems

are anisotropic, to some extent, with respect to permeability, which essentially means that the permeability is larger in some directions than in others. The effect that anisotropy has on the resultant direction of ground water movement is illustrated in Figure 2.3. The resultant flow direction may actually be at an angle other than a right angle to the hydraulic potential contours. In this instance a fractured system with preferred orientations is used to schematically illustrate the effect of anisotropy. It is relatively straightforward to perceive how a lack of continuity in one or more fracture sets or a preferred direction of interconnectivity can affect the relationships between the gradient and the ground water flow direction.

Anisotropy has been recognized as a significant factor in hydrogeologic investigations for many years (e.g., Liakopolus, 1965; Bakthiari, 1971) with recent efforts concentrating on fractured media (e.g., Hsieh, 1983). The work by Hsieh (op. cit.) is particularly relevant, since it is directed at evaluating the quantitative relationships between fractured media and porous media.

Most hydrogeology concepts are derived from field observations of both the surface manifestations of hydrogeologic systems as well as direct measurement of ground water flow systems. Surface manifestations of ground water systems are usually of the form of springs, perennial streams, swamps, vegetation differences etc., while direct measurement of ground water systems is usually accomplished with some form of monitoring system. Ideally a monitoring system or monitoring instrumentation should provide for direct measurement of the physical and hydrogeochemical properties of both porous and fractured media (Figure 2.4). In many instances, it may not be possible to obtain the required information from a single type of instrumentation. These cases are considered later in this section. The most familiar hydrogeologic monitoring instrumentation is the common well (Figure 2.5). Wells are

## Gradient - Flow Direction Relationship

- Anisotropy
- Continuity of individual fractures
- Interconnection of fracture sets

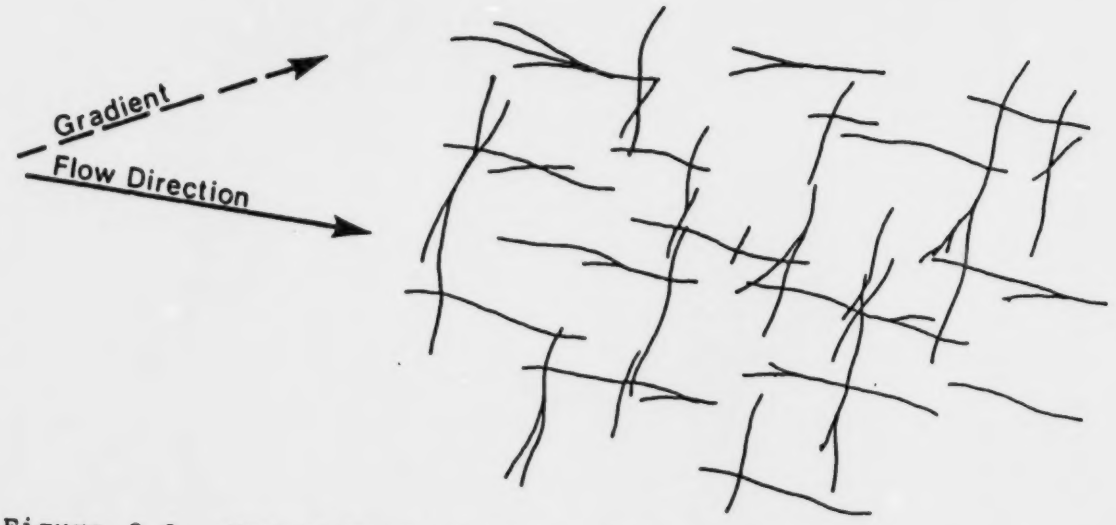



Figure 2.3 Ground Water Flow Direction in a Fractured or Anisotropic Media is Dependent on the Continuity of Individual Fractures, Interconnection of Fracture Sets and Relationship between Gradient Direction and Principal Axes of Permeability

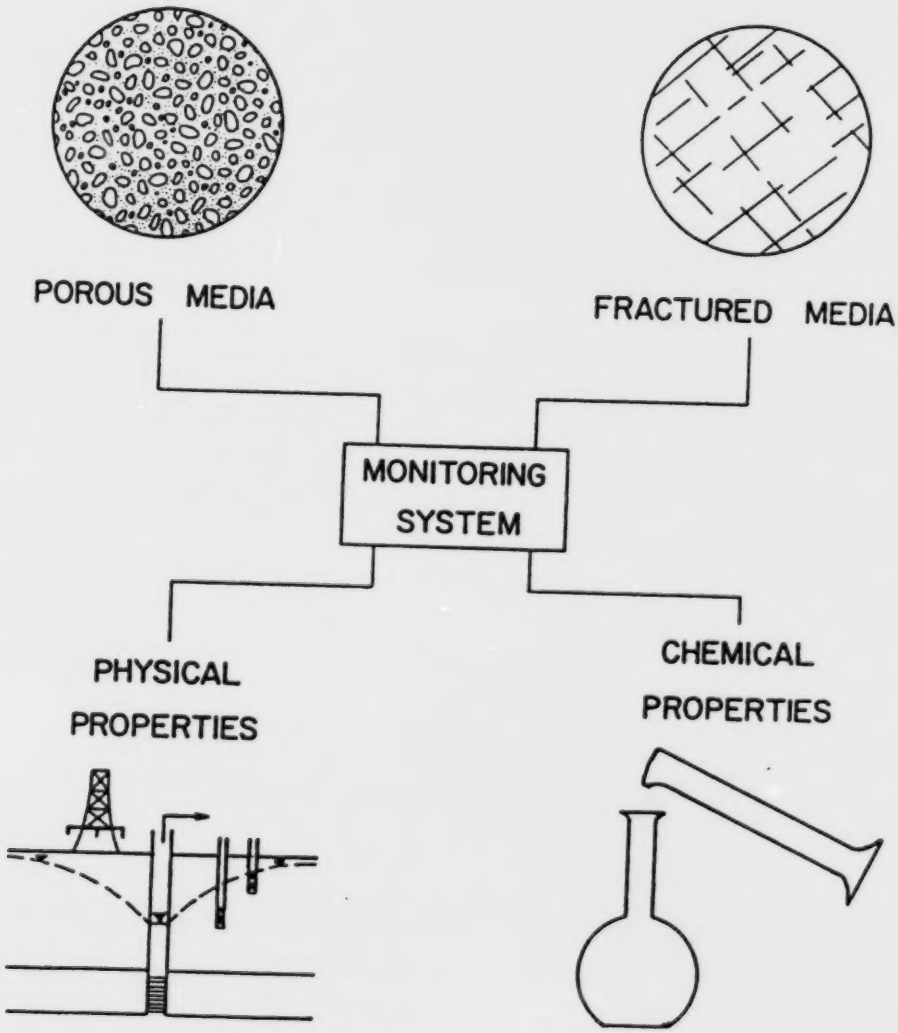



Figure 2.4 Hydrogeologic Monitoring System Requirements

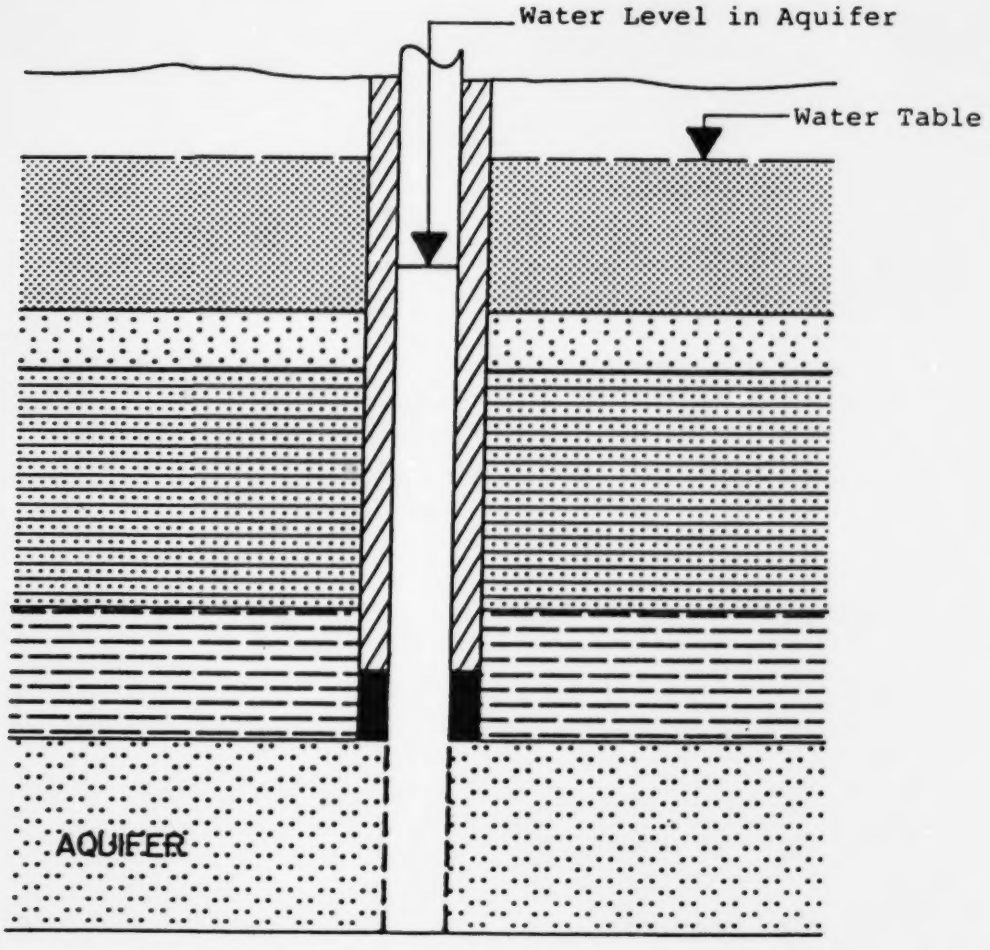



Figure 2.5 Standard Well used for Physical and Chemical Data Gathering

usually installed for the purpose of removing ground water from a hydrogeologic system for domestic, municipal or industrial purposes. A well is usually screened in (or draws water from) an aquifer, which may be defined as a subsurface hydrogeologic unit capable of providing a sufficient quantity of chemically acceptable water for whatever purpose the well is designed.

No matter what type of instrumentation is employed in a hydrogeologic investigation, there is usually an inherent assumption that the monitoring instrument is sampling, for either physical or chemical properties, a "Representative Elementary Volume" (REV) of the hydrogeologic regime. A REV is the smallest volume of porous or fractured medium for which the hydrogeologic processes (e.g., Darcy's law) are valid. Figure 2.6 illustrates the concept of a REV with respect to the sampling scale, or a size of the system that is sampled by a particular type of instrumentation. Physical hydrogeologic properties such as porosity, and chemical properties such as an ionic concentration will vary in magnitude at very small scales (perhaps from pore to pore). However, a sufficiently large sampled volume of the system should provide adequate averaging to consider the measurement representative. A current topic of research interest is whether the REV's for flow, solute transport and geochemical mechanisms are of sufficiently similar size to allow transference of, for instance, the velocity derived from the ground water flow equation to the solute transport equation.

It is clear that the size of the REV in a porous or fractured system depends on the amount and degree of interconnection between pore spaces, which depends on such properties as grain size, compaction, lithification, secondary mineralization, fracture and fracture set orientation, fracture length and continuity, etc. In other words, the geologic and/or structural framework of the media virtually defines the hydrogeologic system's REV.

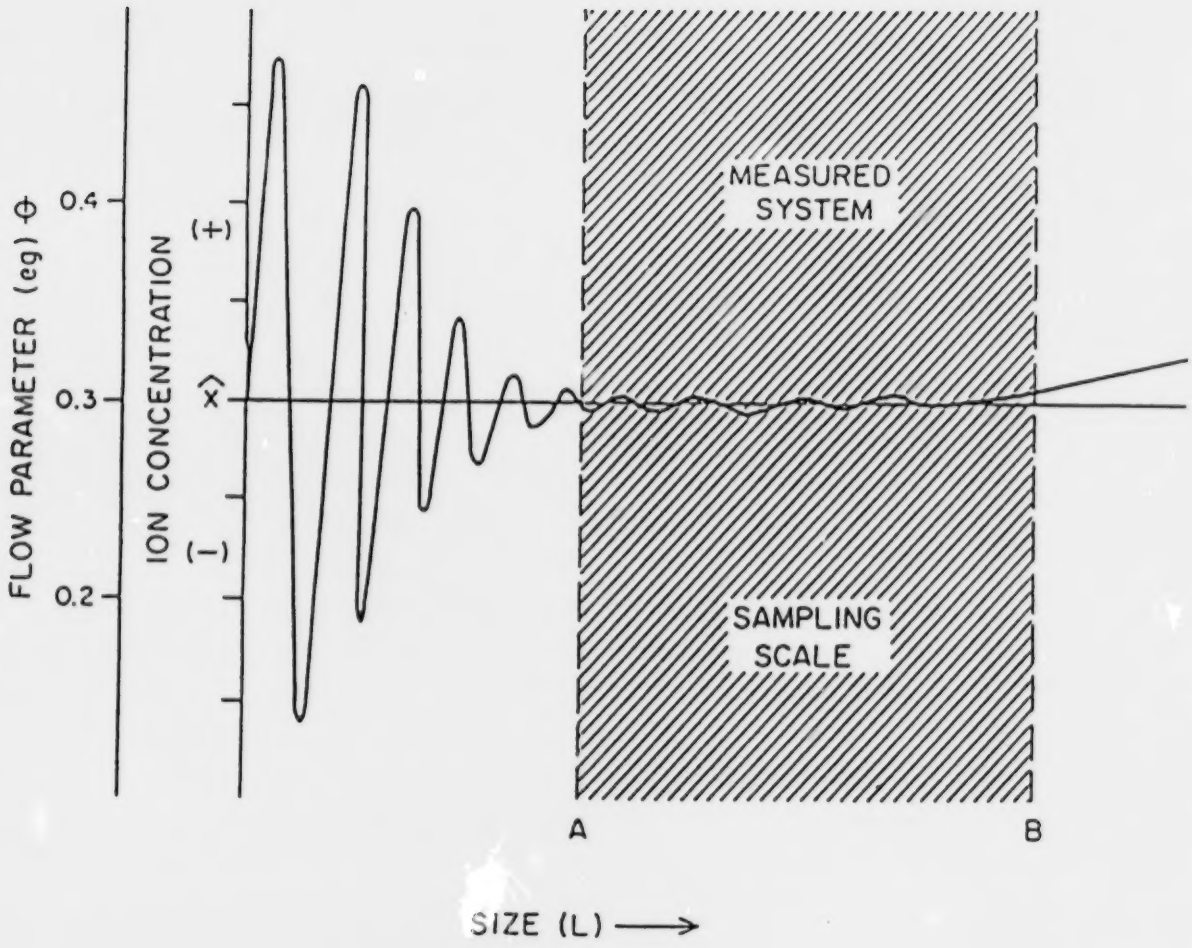



Figure 2.6 Correlation of Measured System Sampling Scale to Variations in Parameter Values (from Grisak et al., 1978)

A second very important aspect to the concept of REV's is the scale of interest. If the scale of interest is that of laboratory column measurements on packed sand cores, the REV is likely of the order of several $cm^3$. If the scale of interest is from the Appalachians to the Hudsons Bay Lowland, or from the Rockies to Lake Winnipeg, the REV may be, for practical purposes, of the order of tens to hundreds of $km^3$. It is difficult to be absolute about the dimensions of a REV in fractured crystalline rock; however, it is clear that the REV is dependent on the fracture spacing, orientation and hydraulic continuity (length), all of which determine the degree of interconnection and the effective bulk hydraulic properties of the formation. At the present time there is no direct manner of relating observed fracture characteristics to subsurface hydrogeologic properties, although national and individual research programs are beginning to develop the data base on which relationships may possibly be developed.

For the two different systems, a porous and a fractured media, shown on Figure 2.7, the same instrument, i.e., a standard sized well or piezometer, may not be suitable as a sampling instrument. (A piezometer is simply a small diameter well, usually installed for the purpose of monitoring a hydrogeologic system rather than for extracting ground water.) Other types of instrumentation have been developed over the years for the purpose of adequately sampling or monitoring hydrogeologic systems. These include multilevel systems in single boreholes (Figure 2.8 and 2.9) for pressure measurement and/or geochemical sampling, as well as multiple piezometer installations as close to one another as possible, known as a piezometer nest (Figure 2.10).

The monitoring instrumentation is usually used to verify or measure the absolute magnitude of the hydraulic nature of a hydrogeologic system which has been conceptualized based on the

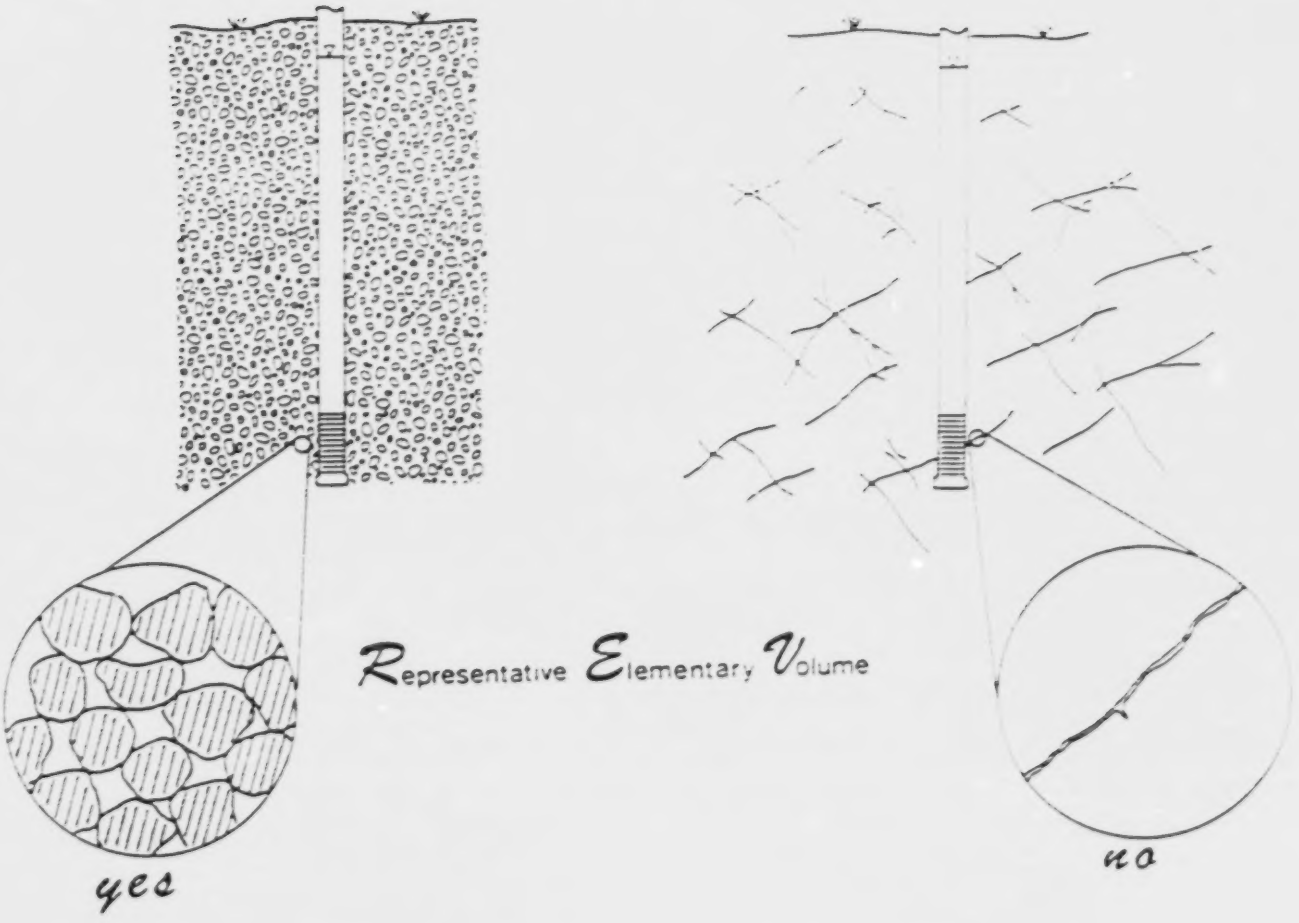



Figure 2.7 Representative Elementary Volume (REV) in the Context of Porous and Fractured Media

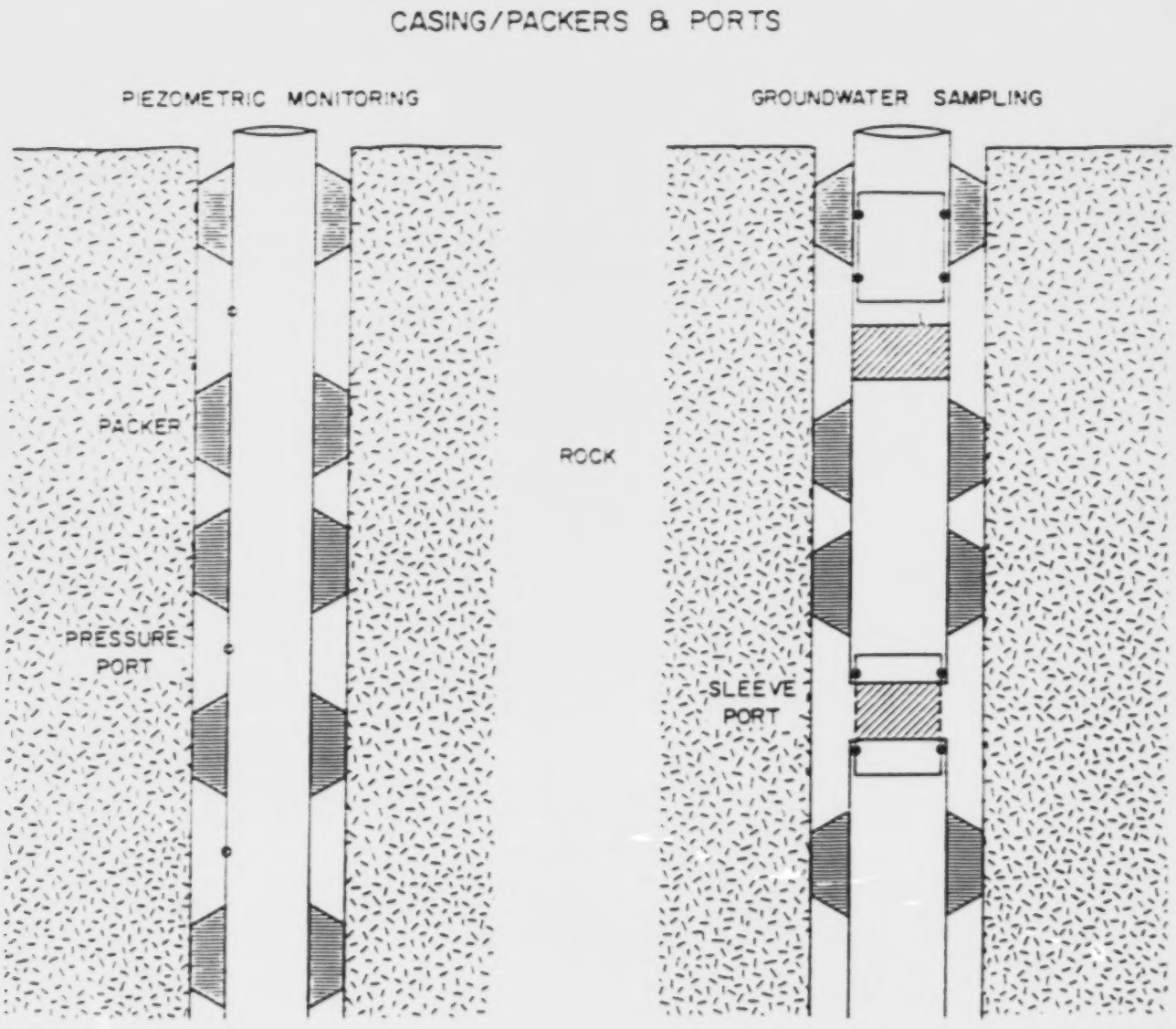
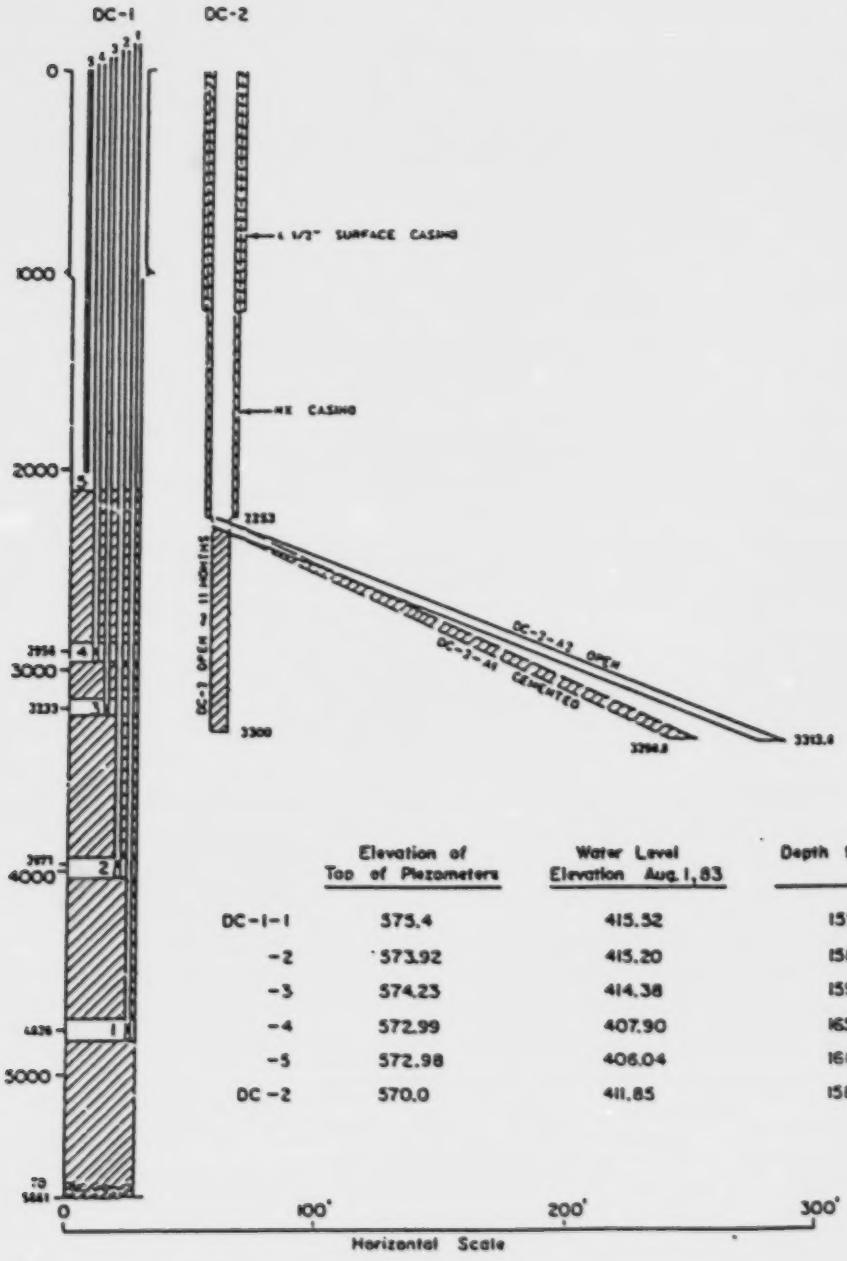



Figure 2.8 Multiple-level Sampling and Monitoring Systems Employed in Recent Years




| | Elevation of Top of Piezometers | Water Level Elevation Aug. 1, 83 | Depth to Water |
|---|---|---|---|
| DC-1-1 | 575.4 | 415.52 | 159.88 |
| -2 | 573.92 | 415.20 | 158.72 |
| -3 | 574.23 | 414.38 | 159.85 |
| -4 | 572.99 | 407.90 | 165.05 |
| -5 | 572.98 | 406.04 | 166.94 |
| DC-2 | 570.0 | 411.85 | 158.15 |

Figure 2.9 Relative elevations and distances of DC-1 piezometers and DC-2 boreholes at Hanford, Washington (from GTC, 1983)

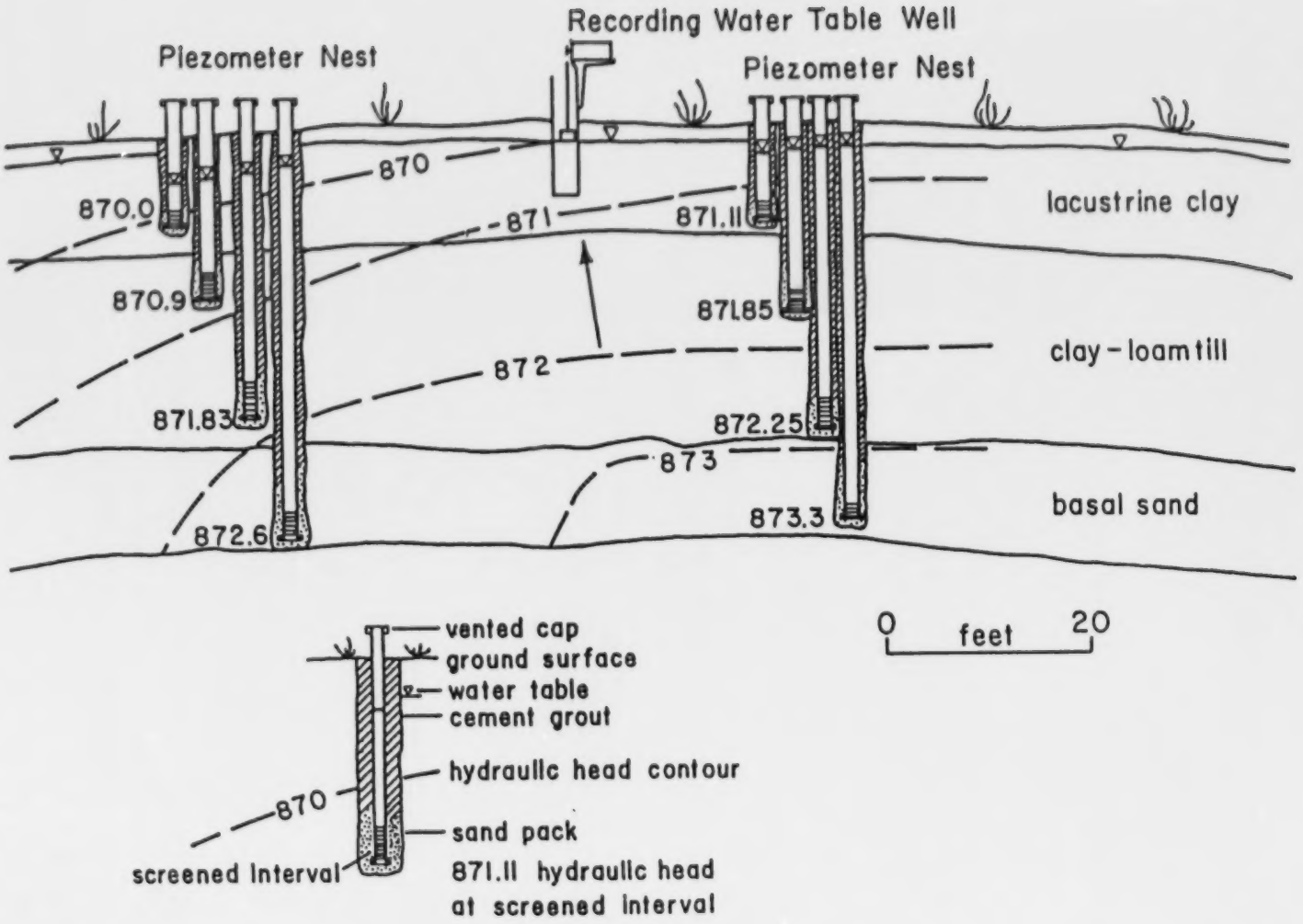



Figure 2.10 Illustration of Piezometer Nests and the Determination of Hydraulic Head Contours from Water Level Measurements (from Grisak and Jackson, 1978)

available geologic, hydrologic and topographic data. One of the most important aspects of conceptualizing a hydrogeologic flow system is to assess the hydrogeologic boundaries (Figure 2.11). The boundary conditions are critical because the material balance within the volume of geologic media is determined by the boundaries.

An example of a three-dimensional hydrogeologic system in a crystalline rock context is schematically illustrated in Figure 2.12. Although the system may appear straightforward, there are a number of aspects which have been considered to develop the conceptualization. These include:

- small scale local flow systems in the near surface; topographically controlled, with flow in surficial deposits or near-surface fractured zone;
- deeper ground water flow generated in the topographic high of the basin;
- topographic high of the basin on the north is a flow divide for all the depths of interest in the hydrogeologic system;
- structural controls on the ground water flow regime allow interbasin flow to the west.

### 2.1.1 Governing Equations

The following equations which govern fluid flow in a saturated medium are presented in terms of pressure, rather than hydraulic head, because the pressure component of the hydraulic head is dependent on the fluid properties. The flow equation describing the transient pressure distribution in the

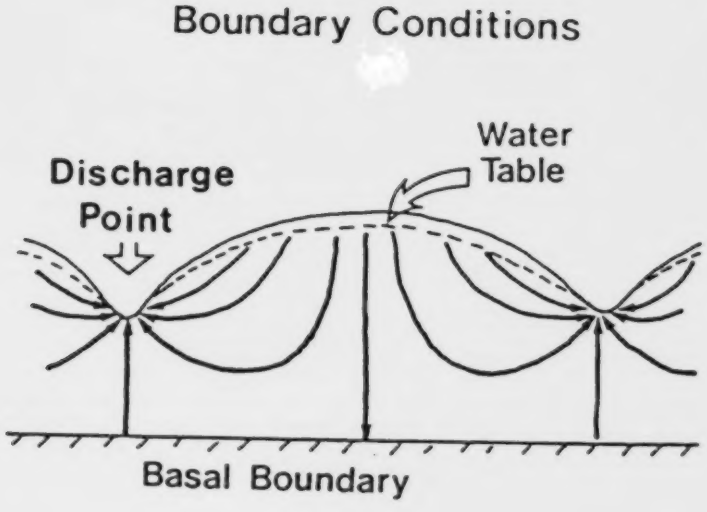



Figure 2.11 Boundary Conditions are Required for Conceptualization of a Hydrodynamic System

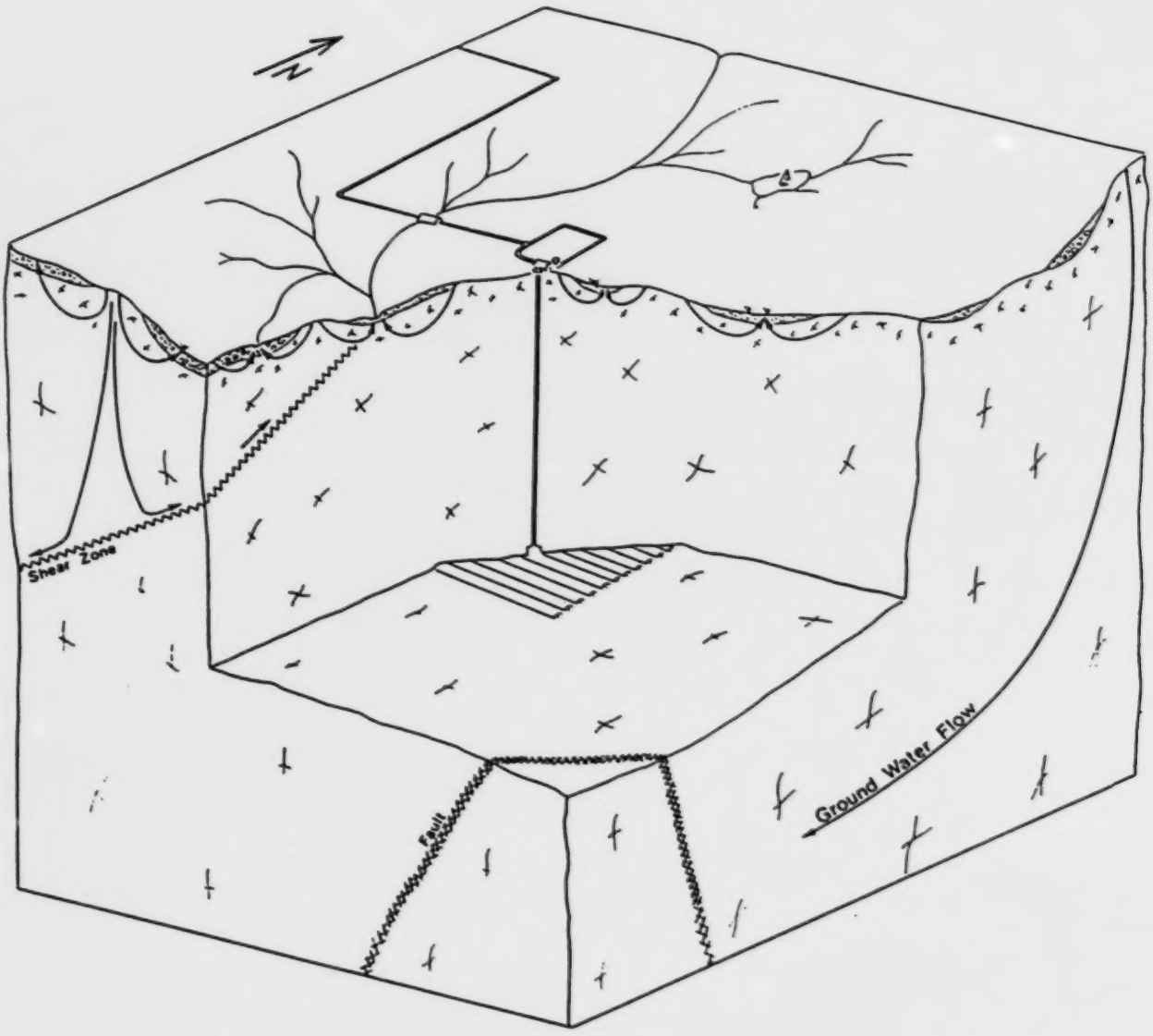



Figure 2.12 Conceptualization of a Repository Situated at Depth in a Topographic and Structurally-Controlled Hydrogeologic System

liquid phase in a saturated porous medium can be written in two dimensions (cross-section)

$$L(p) = \rho (C_R + \theta C_w) \frac{\partial p}{\partial t} + \rho\theta C_T \frac{\partial T}{\partial t}$$

$$- \frac{\partial}{\partial z} \left[ \frac{\rho}{\mu} \left( k_{xx} \frac{\partial p}{\partial x} + k_{xz} \left( \frac{\partial p}{\partial z} + \rho g \right) \right) \right] \qquad (2.1)$$

$$- \frac{\partial}{\partial z} \left[ \frac{\rho}{\mu} \left( k_{zx} \frac{\partial p}{\partial x} + k_{zz} \left( \frac{\partial p}{\partial z} + \rho g \right) \right) \right] + Q^P$$

$$= 0$$

where L is the differential operator defined in the flow region, p is the pressure, $\rho$ is the density of the liquid phase, $C_R$ is the vertical compressibility of the medium, $\theta$ is the porosity, $C_w$ is the compressibility coefficient of the liquid phase, t is time, $C_T$ is the fluid thermal expansivity, x and z are the Cartesian directions with z the vertical, $\mu$ is the dynamic viscosity of the liquid phase, $k_{xx}$, $k_{xz}$, $k_{zx}$ and $k_{zz}$ are the components of the local intrinsic permeability tensor, g is the gravitational constant, and $Q^P$ is the time rate of supply (positive) or extraction (negative) of liquid mass per unit volume at specified locations.

The Darcy velocities, $q_x$ and $q_z$, of the liquid phase in the x and z directions, respectively, are:

$$q_x = -\frac{1}{\mu} \left[ k_{xx} \frac{\partial p}{\partial x} + k_{xz} \left( \frac{\partial p}{\partial z} + \rho g \right) \right] \qquad (2.2a)$$

$$q_z = -\frac{1}{\mu} \left[ k_{zx} \frac{\partial p}{\partial x} + k_{zz} \left( \frac{\partial p}{\partial z} + \rho g \right) \right] \qquad (2.2b)$$

The average interstitial pore water velocities, $V_x$ and $V_z$, are calculated by dividing the Darcy fluxes, $q_x$ and $q_z$, respectively, by the porosity. This is expressed:

$$V_x = \frac{q_x}{\theta} \tag{2.3a}$$

$$V_z = \frac{q_z}{\theta} \tag{2.3b}$$

When the principal components of the intrinsic permeability tensor are measured in directions different from the x-z Cartesian system (with z the vertical), they can be formulated in the Cartesian system using the equations given by Bear (1972, p. 141). The components, $k_{xx}$, $k_{xz}$, $k_{zx}$ and $k_{zz}$ in the Cartesian system are:

$$k_{xx} = \left(\frac{1}{2}\right)\left[k_{xx}' + k_{zz}' + (k_{xx}' - k_{zz}') \cos 2\phi\right] \tag{2.4a}$$

$$k_{zz} = \left(\frac{1}{2}\right)\left[k_{xx}' + k_{zz}' - (k_{xx}' - k_{zz}') \cos 2\phi\right] \tag{2.4b}$$

$$k_{xz} = k_{zx} = -\left(\frac{1}{2}\right)(k_{xx}' - k_{zz}') \sin 2\phi \tag{2.4c}$$

where $\phi$ is the angle between the x-axes of the two coordinate systems, and $k_{xx}'$ and $k_{zz}'$ are the principal instrinsic permeability components in the measured coordinate system.

The initial conditions necessary for the solution of Equation (2.1) are:

$$p\,(x,z,0) = \tilde{p}_o\,(x,z) \tag{2.5}$$

where $\tilde{p}_o$ are initial prescribed pressures. The Dirichlet or prescribed pressure boundary conditions can be expressed as:

$$p\,(x,z,t) = \tilde{p}\,(x,z,t) \quad \text{on } \Gamma_1 \tag{2.6a}$$

and the Neumann or flux boundary conditions can be expressed as:

$$\left\{\frac{1}{\mu}\left[(k_{xx} + k_{zx})\frac{\partial p}{\partial x} + (k_{xz} + k_{zz})\left(\frac{\partial p}{\partial z} + \rho g\right)\right]\right\}\bar{n}$$

$$= -q_n^P \qquad \text{on } \Gamma_2 \qquad (2.6b)$$

where $\tilde{p}$ are prescribed pressures, $\Gamma_1 + \Gamma_2 = \Gamma$, the total boundary of the region, $q_n^P$ is the prescribed liquid flux across a Neumann boundary and $\bar{n}$ is the unit outward normal vector to the boundary.

### 2.1.2 Research Techniques and Parameter Evaluation (Data Required to Quantitatively Evaluate The Ground Water Flow Regime)

#### 2.1.2.1 Hydrogeologic Boundary Conditions

The hydrogeologic domain must be bounded with either flux, pressure (or head) or permeability boundaries. These are generally inferred from the geometry and geologic conditions combined with the measured hydraulic heads and permeabilities. Major structural features such as faults are often considered as boundaries due to their likely permeability contrasts with the surrounding media.

#### 2.1.2.2 Permeability

The permeability distribution within a hydrogeologic system is required to develop a conceptualization and quantification of the flow/velocity field. The three-dimensional permeability ellipsoid is required in order to calculate the resultant flow directions. Laboratory permeability testing

methods include the use of permeameter tests on cores and permeability estimates from grain-size distributions. Field permeability tests using one or more boreholes, wells or piezometers as instrumentation are conducted to provide spatial estimates of permeability to assist in the overall characterization of the hydrogeologic system. A brief review of field permeability testing methods for various permeability ranges is presented in Table 1.1. Some tests are better suited than others to a particular field situation depending on such factors as magnitude of permeability, length of test and scale of the geologic system being tested. Figure 2.13 illustrates the range of permeabilities over which the different types of tests are most suitable. These field tests allow assessments, in some cases, of the continuity of various geologic or hydrostratigraphic units or zones.

Permeability testing is conducted in several ways; however, most testing and analysis methods are concerned with a pressure or hydraulic head change in a test interval and the return of the pressure in the interval to the pretest pressure conditions. The hydraulic parameters of the formation are related to the duration and shape of the recovery curve. The pretest pressure in the interval is related to the pressure history to which the interval has been subjected prior to testing.

There are two principal types of permeability testing situations; a) wells and piezometers and b) packer tests. In order for any well or piezometer to provide meaningful hydraulic head data, the following criteria must be satisfied:

- the piezometer must be hydraulically connected to the zone of interest, installed so as to minimize the time lag (Hvorslev, 1951) for representative measurements;

TABLE 1.1 Field Permeability Tests

| TEST | DESCRIPTION | APPLICATION | REFERENCE |
|---|---|---|---|
| Pumping; injection | Pumping or injection to a well at a constant or step-wise varying rate. Monitor head response in the pumping/ injection well. Usually monitor head response in observation wells. | Rocks of high permeability | Theis (1935); Bear (1979) |
| Slug injection; slug withdrawal | Single-well test. Monitor head (water level) increase or decrease with time in response to an induced sudden change in water level in the well. | Rocks of high permeability | Cooper et al. (1967) |
| Pulse injection; pulse withdrawal | Single-well test. Monitor pressure or head increase or decrease with time in response to an induced sudden change in head. The test interval is isolated from the remainder of the borehole or well by packers. | Rocks of low permeability | Bredehoeft and Papadopulos (1980) |
| Drill stem; pressure buildup | Single-well test with the test interval isolated by packers. There is a "flow period" during which formation fluid from the test interval flows into the tubing string above it. The interval is then shut in for a pressure "build-up period". | Rocks of high permeability | Matthews and Russell (1967); Earlougher (1977); Lee (1982) |
| Cross-hole interference | Multiple-well test. The pressure or head in one well or packer-isolated interval is varied suddenly or in a recurrent (eg. sinusoidal) manner and the pressure in zones in other monitoring wells is monitored. | Rocks of high permeability | Earlougher (1977); Black and Kipp (1981) |

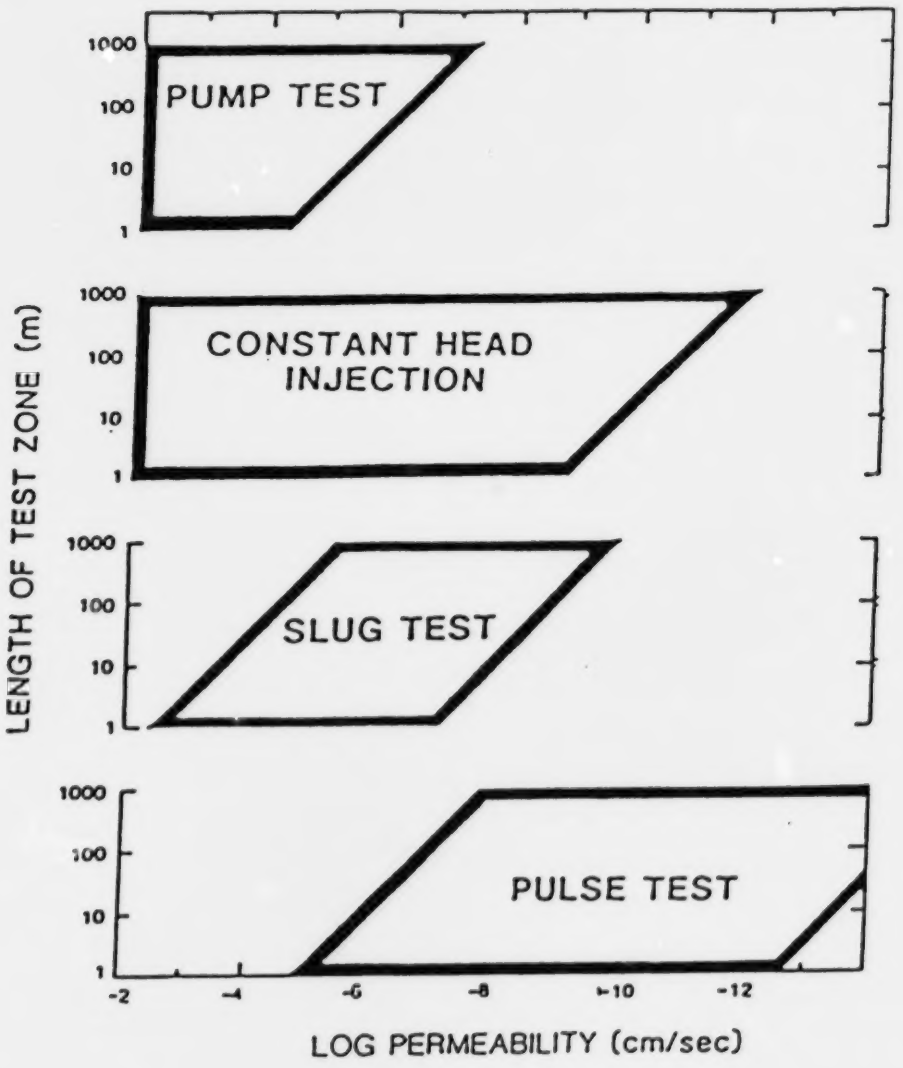



Figure 2.13 Guide to Selection of Hydraulic Test Techniques for Permeability Measurement (Wilson et al., 1979)

- the piezometer intake zone (screen) must be hydraulically isolated from other zones with potentially different hydraulic heads;

- the standpipe (tubing) portion of the piezometer must be free from leaks.

The time lag referred to above is simply the time difference between a change in hydraulic head in the formation and the manifestation of that change in the water level measured in the piezometer. The "Basic Time Lag" $T_0$ (Hvorslev, 1951) is defined as the time that would be required for complete equalization of the head difference between the formation and the piezometer if the original rate of inflow (or outflow) to the piezometer caused by that head difference were maintained. In very impermeable formations $T_o$ is of the order of several months.

Hydraulic isolation of individual zones is usually accomplished by cement or grout seals (see Figures 2.9 and 2.10). The cement seal in a piezometer or well installation must provide for good bonding to casing and formation and must, at least, be less permeable than the formation between the isolated intervals; otherwise a certain amount of 'short circuiting' is inevitable. Most common downhole cement mixtures are generally considered to have a hydraulic conductivity of the order of $10^{-8}$ to $10^{-9}$ m/s or less after a set period of about 10 days (Smith, 1976; Roy et al., 1979).

Packer tests refer to testing in open boreholes where isolation of an interval is provided by means of expandable packers lowered in the borehole and expanded against the borehole wall. Packer tests are usually conducted in either a single or double packer configuration. Recent experimental

packer testing systems employ multiple 'guard' packers above and below the interval (Peterson et al., 1983).

A recently designed 'umbilical' through-the-bit (TTB) single packer testing configuration (GTC, 1982) is illustrated in Figures 2.14 and 2.15. Figure 2.14 is a schematic representation of the overall TTB equipment configuration. The downhole components are lowered on bundled cable consisting of air/fluid lines and a multiconductor geophysical cable which also serves as the strain member. A fluid rotary is used to regulate pressure in the fluid/air lines to prevent collapse due to hydrostatic pressures.

The downhole components of the TTB system are detailed in Figure 2.15. The single packer extends through the drill bit and inflates pneumatically (or with hydraulic pressure) sealing the lower portion of the borehole. There are pressure and temperature measurements within the test interval and within the drill stem. A piston for transient pulse tests is located within the test interval, which allows for injection testing. Umbilical single packer systems similar to that illustrated in Figures 2.14 and 2.15 have been employed in numerous field, investigations, particularly in the Canadian Nuclear Fuel Waste Management Program.

Double packer systems are used by numerous research programs. In most instances the double packer systems are configured around drill tubing as both the backbone of the system and the fluid carrier for hydraulic (i.e., permeability) testing. The Lynes double packer system, in common use in several research programs, is schematically illustrated in Figures 2.16 and 2.17.

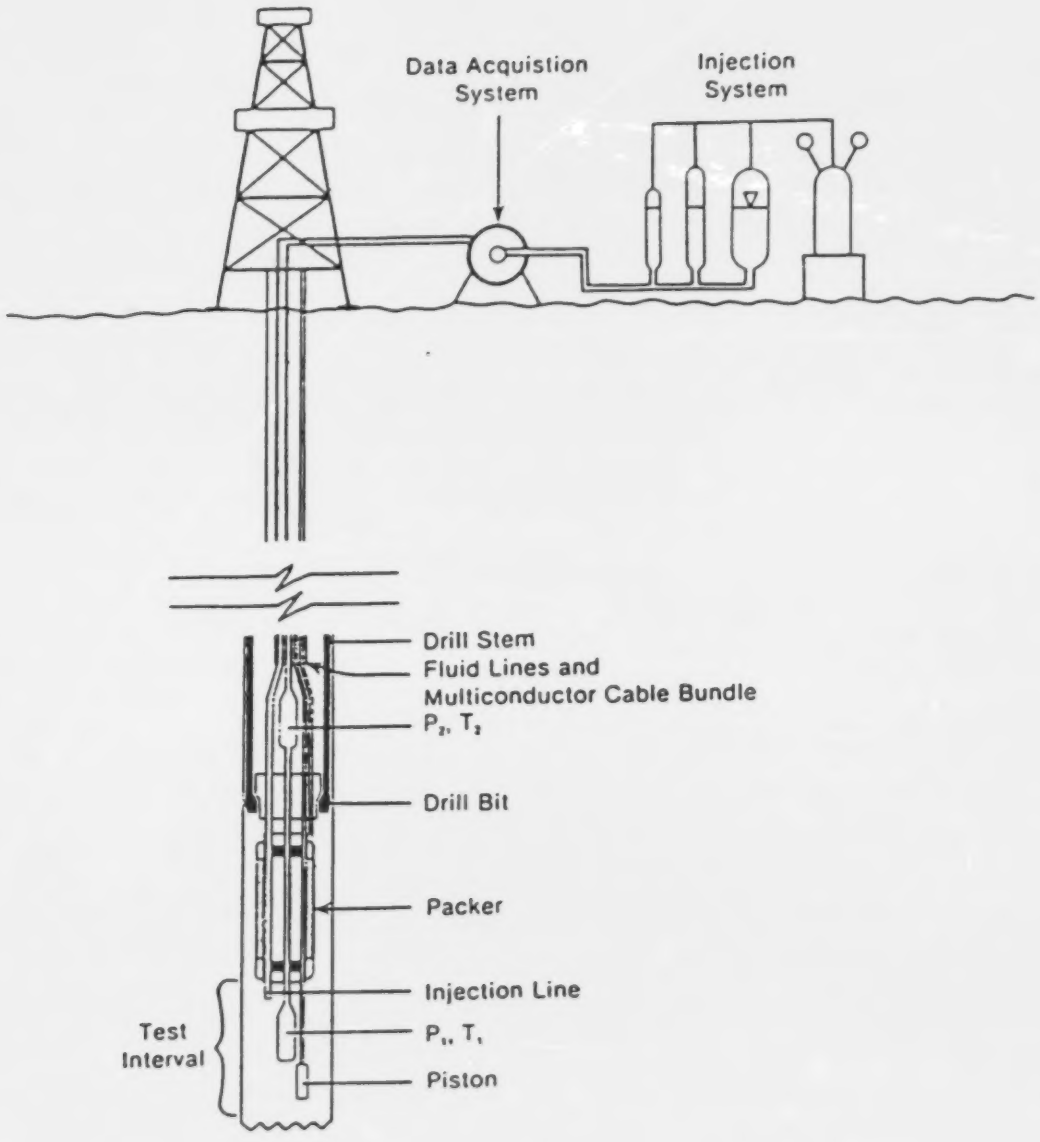



Figure 2.14 Schematic Diagram Illustrating Complete Single Packer Through-the-Bit (TTB) Equipment Configuration (from GTC, 1982)

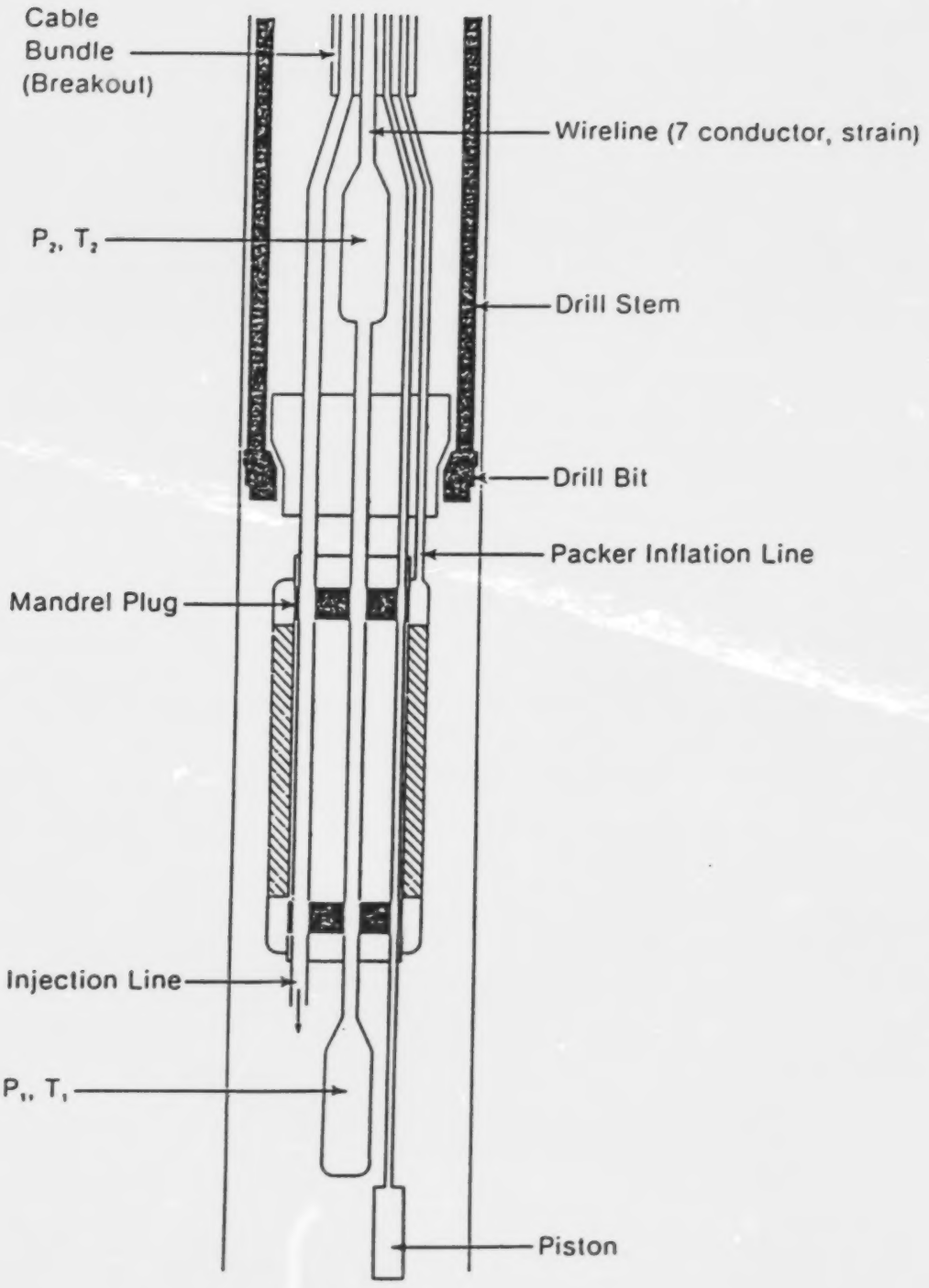



Figure 2.15 Schematic Diagram Illustrating Downhole Detail of Single Packer TTB Equipment Configuration (from GTC, 1982)

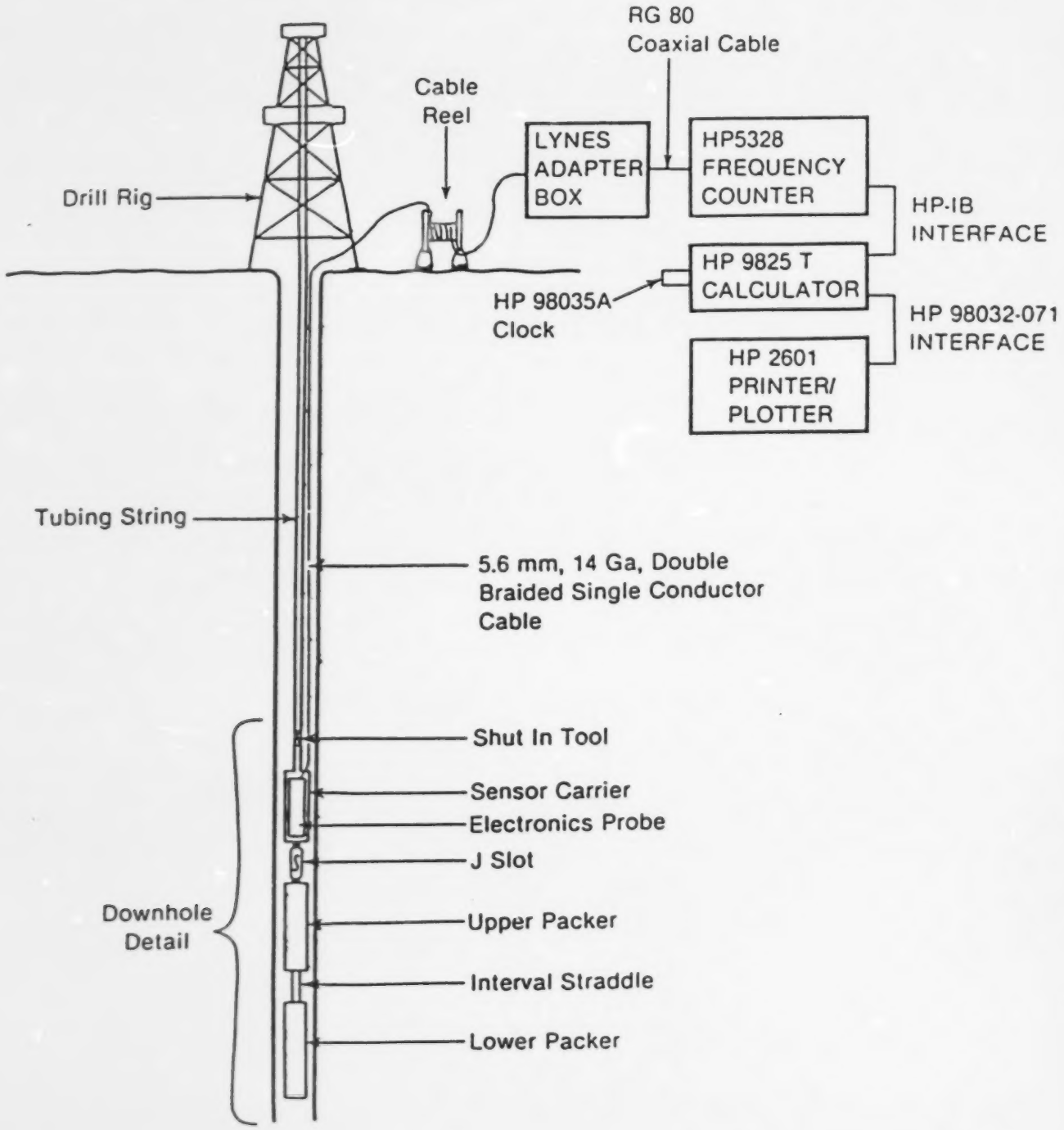



Figure 2.16 Schematic Diagram of Complete Double Packer Hydraulic Testing Equipment based on Lynes Inc. Equipment (from GTC, 1982)

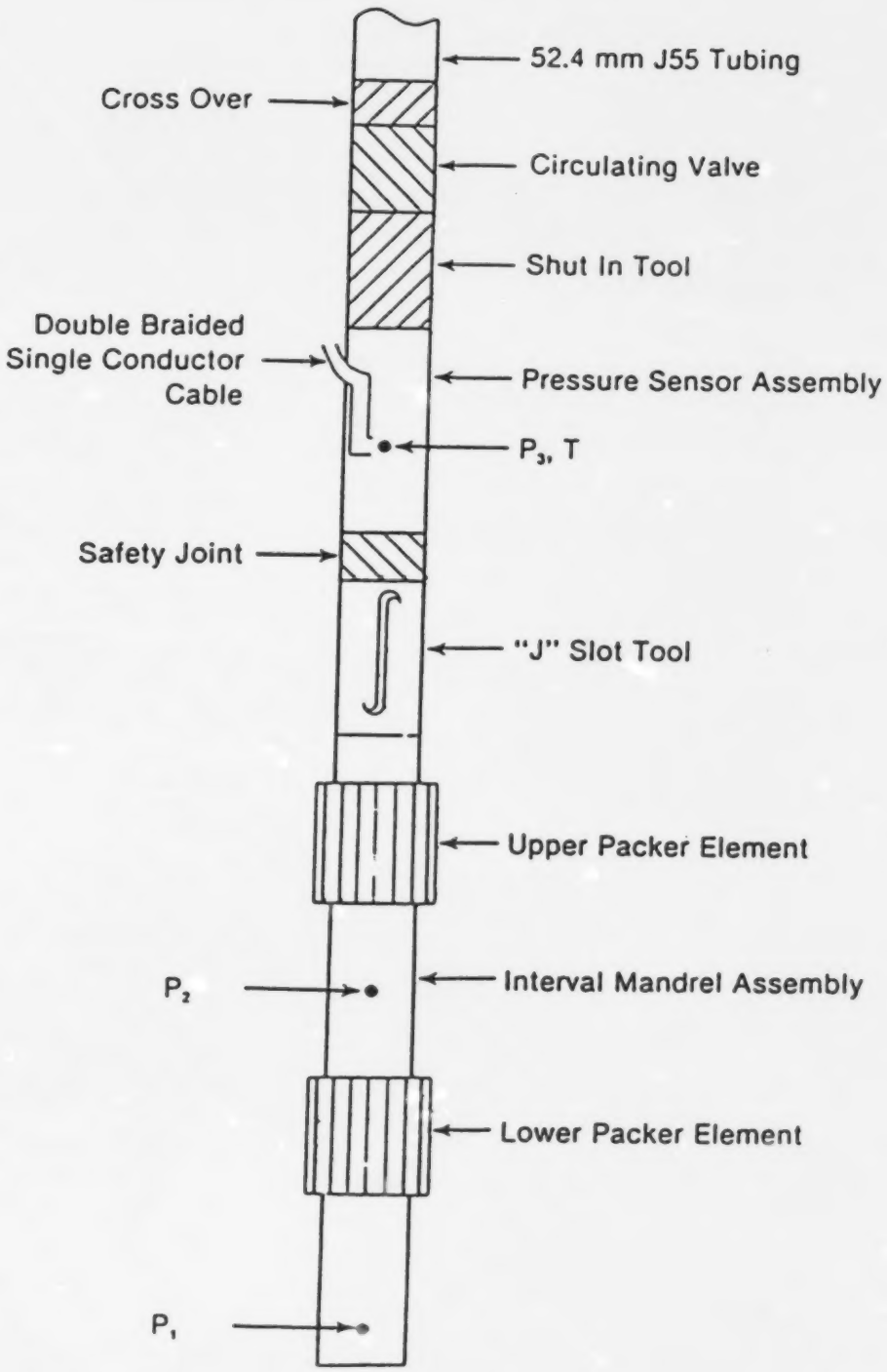



Figure 2.17 Schematic Illustration of Downhole Double Packer Assembly (Lynes Inc. Equipment) (from GTC, 1982)

The Lynes double packer system consists of a straddle packer assembly interfaced with a downhole electronics/sensor carrier which transmits data through a geophysical logging cable to surface monitoring readout/data storage equipment. Pressure data for the zones above and below, as well as in the straddled interval, are provided by porting the three transducers to these zones. Incorporated above the entire assembly is a shut-in tool which isolates the straddled interval from the open tubing string above the zone of interest. This isolation capability is critical for performing static as well as transient tests in the borehole.

The quoted accuracy and resolution of the quartz transducers are 0.05 percent and 0.005 percent of full scale, respectively. Temperature accuracy and resolution are quoted as 1°C and 0.1°C over the range 0°C to 100°C. Temperature readings are those at the probe locations, not in the intervals above, below, and within the packers. The pressure and temperature data are transmitted from the downhole sensor probe over a single conductor wireline to the Lynes interface unit. These downhole data are converted from frequency signals to pressure and temperature readings that are printed, plotted, and stored on magnetic tape.

#### 2.1.2.3 Hydraulic Head or In-Situ Pressure

The hydraulic head distribution is necessary for quantifying the hydraulic gradient and flow directions within a hydrogeologic system. Open piezometers or wells are suitable to obtain hydraulic head measurements in shallow systems. For the investigation of deeper (e.g., 100's to 1000's of metres) flow systems, it is common to measure and report the in-situ pressure using packer systems as described above. If heads are reported, the corresponding reference fluid properties must be stated, due to the pressure and temperature dependence of the properties of water.

2.1.2.4 Specific Storage
Medium Compressibility
Fluid Compressibility

The specific storage of a saturated geologic medium is defined as the volume of water that a unit volume of the medium releases from storage under a unit decline in hydraulic head. This volume of water released from storage is dependent upon the compressibility of the fluid and the medium. Specific storage $S_s$ can be expressed

$$S_s = \rho g \, (C_R + \theta C_W) \qquad (2.7)$$

where $\rho$ = fluid density
$g$ = gravitational constant
$C_R$ = rock or medium compressibility
$\theta$ = porosity
$C_W$ = fluid compressibility

The mathematical representation of the specific storage is dependent on the definition of the fluid and medium compressibilities. A detailed review of the various conventions for defining the medium compressibility is given by Narasimhan and Kanehiro (1980). The expression for specific storage noted above is based on a medium compressibility definition utilizing a normalization with respect to the bulk volume of the sample being tested. The medium compressibility $C_R$ is expressed

$$C_R = \frac{\Delta\varepsilon}{\Delta\sigma} = -\frac{1}{V_b} \frac{\Delta V_v}{\Delta \sigma} \qquad (2.8)$$

where $V_b$ = bulk volume
$V_v$ = volume of voids
$\Delta\varepsilon$ = change in strain
$\Delta\sigma$ = change in effective stress

If drainage is allowed, then $\Delta\sigma = \Delta p$ where p is the fluid pressure. This definition for $C_R$ is referred to as the coefficient of volume change in the soil mechanics literature. For uniaxial loading, the ratio of stress to strain is the Young's modulus. For three-dimensional loading, it is necessary to consider the bulk modulus which accounts for the volumetric strain. For the special case of equal stress in all directions, the bulk modulus B is expressed:

$$B = \frac{\sigma}{\Delta V_b / V_b} = \frac{E}{3\ (1-2\ u)} \qquad (2.9)$$

where E = Young's modulus
u = Poisson's ratio

The medium compressibility is the reciprocal of the bulk modulus for three-dimensional cases:

$$C_R = \frac{3\ (1-2\ u)}{E} \qquad (2.10)$$

The stress-strain relationships for rocks or porous media are usually measured under laboratory conditions using a variety of experimental techniques. The methodologies involve measurements of load and corresponding strain under uniaxial or triaxial conditions.

### 2.1.3 Hydraulic Testing

Hydraulic testing is the term used to describe tests (primarily field tests) which yield the previously discussed parameters relating to ground water movement. In the past few years, considerable hydraulic testing research has been directed towards low permeability media; in particular towards fractured rock. This is due to the selection of potential radioactive disposal sites at depth in rock types such as layered basalts,

tuffs and crystalline rock. Due to the interest and relevance of such research, the following section will consider in detail the principles and difficulties of hydraulic testing in fractured low permeability media. Most of these testing programs are conducted during drilling or immediately thereafter, with long term multilevel installations for longer term hydraulic head measurements. The borehole measurements are usually conducted using single or double packer testing techniques. One of the most critical difficulties in hydraulic testing is to establish the fact that values for the formation parameters such as hydraulic head and permeability are 'representative' of the formation. Since the hydraulic regime in the formation has been disrupted to some extent by the drilling process, these transient effects must be considered in the subsequent testing and interpretation. The following discussion of these effects is precised from GTC (1984). The discussion is directed at the effects on testing of i) pretest borehole pressure history, ii) thermally induced borehole pressure responses and iii) equipment compliance.

For borehole pressure measurements using single or double packer equipment, the monitored pressure response is dependent on a number of parameters including fluid compressibility, fluid density, fluid thermal expansion coefficient, fluid viscosity, fluid heat capacity, formation hydraulic conductivity, formation thermal conductivity, formation heat capacity and, in some cases, their dependence on pressure and temperature. The relative importance of the pressure and temperature dependence of the fluid parameters (see Figure 2.18) is summarized on the following page.

Figure 2.18 Fluid properties relevant to pressure measurement at depth in geologic formations (from GTC, 1983)

Relative Dependence of Borehole and Formation Fluid Parameters on Pressure and Temperature

| | Density | Compressibility | Coefficient of thermal expansion | Viscosity |
|---|---|---|---|---|
| Temperature | high | high | very high | very high |
| Pressure | moderate | high | low | low |

The above dependencies are most relevant to the case of a 'shut-in' interval, that is, when the pressure measurement does not rely on an equilibrated column of water (i.e., piezometer standpipe) for hydraulic head measurement, but rather the pressure of the interval is measured directly with a downhole pressure transducer. Pressure recovery is faster because it is not necessary for the column of water to come to equilibrium with the formation.

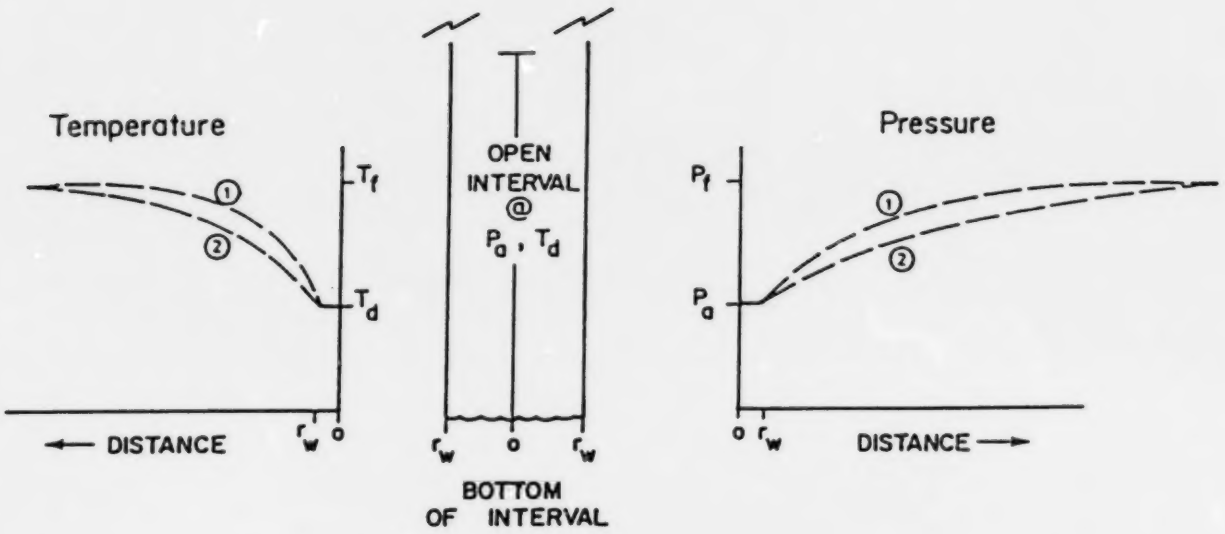
#### 2.1.3.1 Pretest Borehole Pressure History

It is instructive to approach the problem of formation responses to drilling and testing by considering the sequential occurrence of events from the time a geologic interval is intersected during drilling until the completion of a single or double packer test. A pretest pressure condition whereby the in-situ formation pressure is greater than the annulus fluid pressure is employed for illustration purposes. Annulus pressure refers to the downhole absolute pressure corresponding to overflow of the drilling fluid at the wellhead.

Prior to drilling into an interval the pressure of the interval is considered to be at in-situ or formation conditions ($P_f$). As the interval is drilled, the rock is subjected to annulus pressure ($P_a$). A pressure (P) profile is developed radially into the formation in response to the differential pressures between the borehole fluid and the rock. The P-profile into the rock for the test interval is schematically illustrated in Figure 2.19. (In reality, the P-profile will be different between the top and the bottom of the interval, since the upper portion has been open to drilling conditions for a longer period of time than the lower portion). The longer the interval is open the further into the formation the P-profile develops (see Figure 2.19).

The net effect of the "pressure skin" developed during the drilling and pretest period is the establishment of testing conditions that do not meet the assumptions involved in type-curve analysis methods based on analytical solutions of the radial flow equation. For instance, the pressure recovery in the interval will not be simply related to the recovery to in-situ formation pressure; it will also be related to the pre-test pressure history which has created the "skin". The analysis of packer tests using type-curve methods could therefore yield incorrect permeability estimates under conditions where the annulus pressure and formation pressure are different.

In conducting hydraulic tests for obtaining pressure response data for hydraulic conductivity determination, it is necessary to choose testing procedures that take into account the magnitude of the annulus and formation pressures. Schematic pressure response curves for pulse or slug tests are shown in Figure 2.20 for recommended test procedures. For the illustrated injection and withdrawal tests, a strong pressure pulse or head differential is required such that at




$P_a$ annulus pressure

$P_f$ formation (in-situ) pressure

$T_f$ formation (in-situ) temperature

$T_d$ drill fluid temperature

① profile on completion of drilling

② profile after some period following completion of drilling

$r_w$ radius of borehole

Figure 2.19 Schematic pressure (P) and temperature (T) profiles in the rock on completion of drilling an interval (from GTC, 1983)

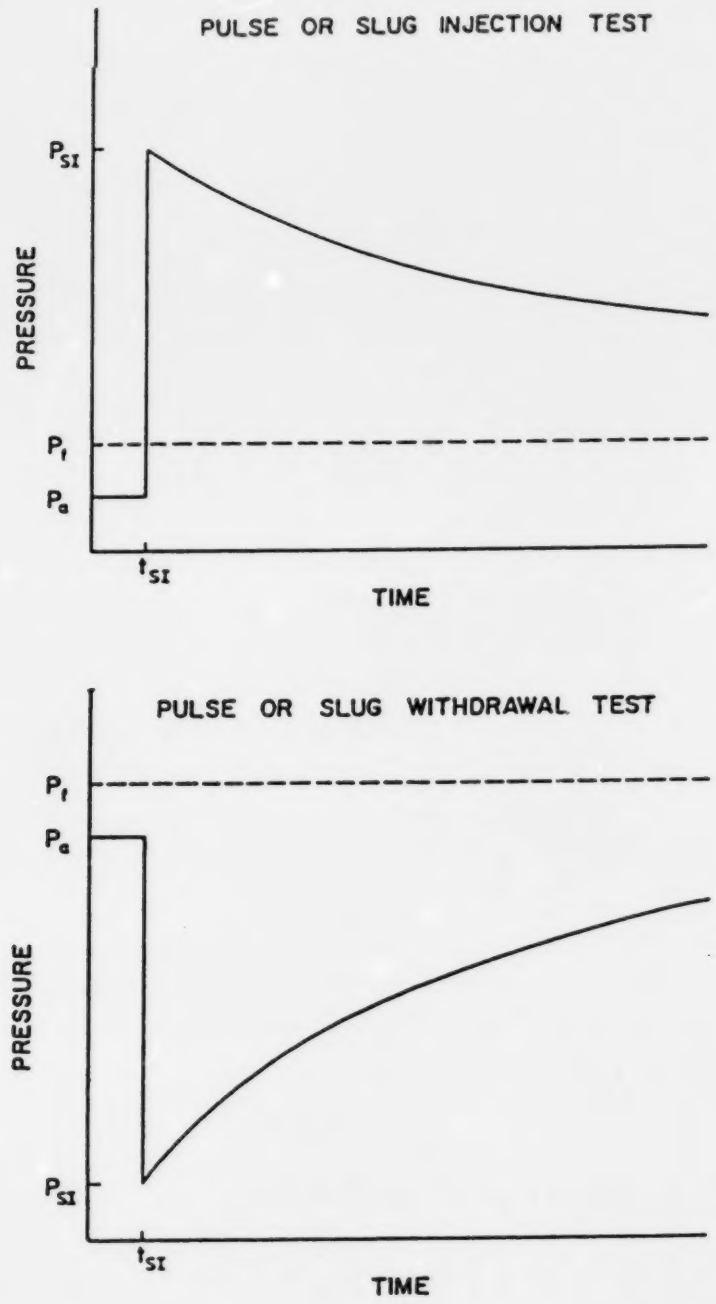



Figure 2.20 Schematic pressure responses for the test interval utilizing adequate test procedures to allow hydraulic conductivity determination (i.e., the pressure $P_{SI}$ at the start of the test is substantially different from annulus pressure $P_a$ or formation pressure $P_f$)

the start of the test, the difference between the pressure induced by either fluid injection or fluid withdrawal ($P_{SI}$), and both annulus pressure ($P_a$) and formation pressure ($P_f$) is much larger than the pressure difference between $P_a$ and $P_f$.

Schematic pressure response curves for pulse or slug tests which would not provide suitable data for correct hydraulic conductivity determinations are illustrated in Figure 2.21. In these cases, a strong pressure pulse or head differential has not been utilized as the starting condition for the tests. The pressure at the start of the test is either equal to the annulus pressure or is not substantially different from the annulus or formation pressure. The net effect of these inadequate test procedures is a high degree of uncertainty in the interpreted hydraulic conductivity. For example, the analysis using type-curve methods of the test where the starting pressure ($P_{SI}$) equals the annulus pressure ($P_a$) could yield a hydraulic conductivity estimate that is several orders of magnitude too low.

A long period open to annulus pressure conditions causes a sufficiently large pressure skin to develop such that, for the purposes of hydraulic conductivity determination, the annulus pressure can be assumed as the "reference" or pretest pressure. With a long borehole open period and adequate testing procedures, the choice of annulus pressure as the reference pressure is preferable to an estimate of $P_f$, since $P_a$ will provide the primary influence on the pressure recovery. In this case the actual formation pressure has a relatively small effect on the measured borehole pressure response during testing. The recovery from the pressure pulse responds as if the annulus pressure were the actual formation pressure.

Borehole pressure history can also be important during a testing sequence involving more than one test (eg., a pulse

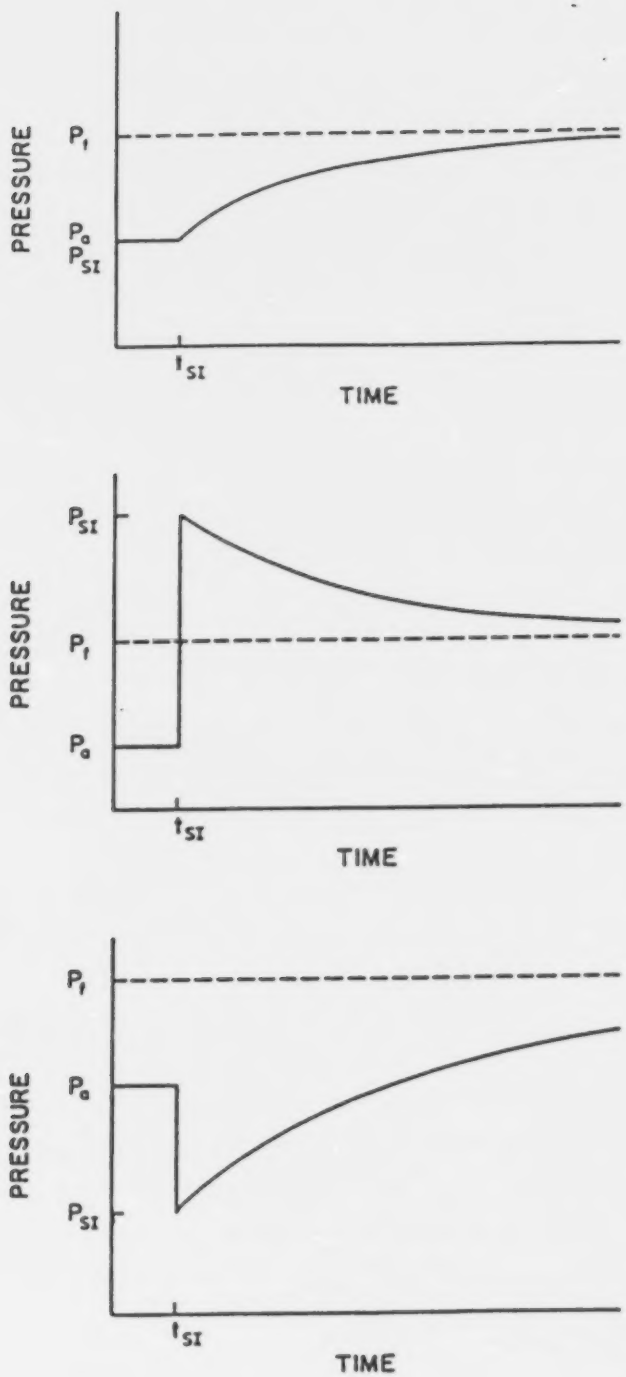



Figure 2.21 Schematic pressure responses for the test interval utilizing inadequate test procedures for hydraulic conductivity determination (i.e., the pressure $P_{SI}$ at the start of the test is at or close to the annulus pressure $P_a$ or formation pressure $P_f$)

withdrawal test followed by a pulse injection test). Each individual hydraulic test causes some modification of the pressure profile radially from the borehole. This pressure skin can have an important effect on the measured borehole pressure response during subsequent tests.

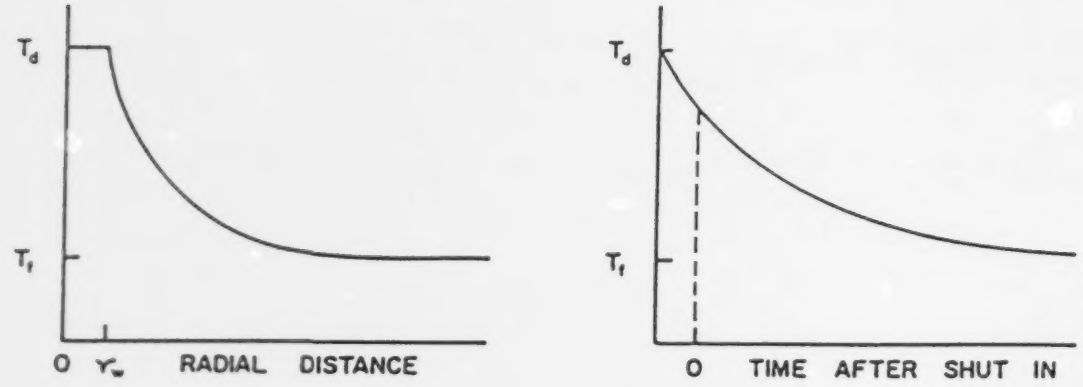
In summary, hydraulic tests should be designed and conducted to minimize the effects of pretest borehole pressure history. When using type-curve analysis methods, the choice of the reference pressure should not be determined solely by type-curve extrapolation but rather should be chosen based on an evaluation of the annulus pressure and estimated formation pressure as well as the length of time that the interval has been exposed to annulus pressure conditions.

2.1.3.2 Thermally-Induced Borehole Pressure Response

Pressure changes resulting from thermal expansion (or contraction) of the borehole fluid can exert a very important effect on the measured borehole pressure response during hydraulic testing. Prior to drilling into an interval the temperature of the interval is considered to be at in-situ or formation conditions ($T_f$). As the interval is drilled, the rock is subjected to drilling fluid temperatures ($T_d$). A temperature profile is developed into the formation in response to the differential temperatures between the borehole fluid and the rock. The temperature profile into the rock is illustrated schematically in Figure 2.19. The longer the interval is open the further into the formation the temperature profile develops.

Temperature profiles for cases with the drill fluid temperature greater than the formation temperature and also less than the formation temperature are shown schematically in Figure 2.22. The net effect of the development of these temperature profiles in the formation is a variation in the borehole fluid temperature during the hydraulic testing

(a) DRILL FLUID TEMPERATURE > FORMATION TEMPERATURE




(b) DRILL FLUID TEMPERATURE < FORMATION TEMPERATURE

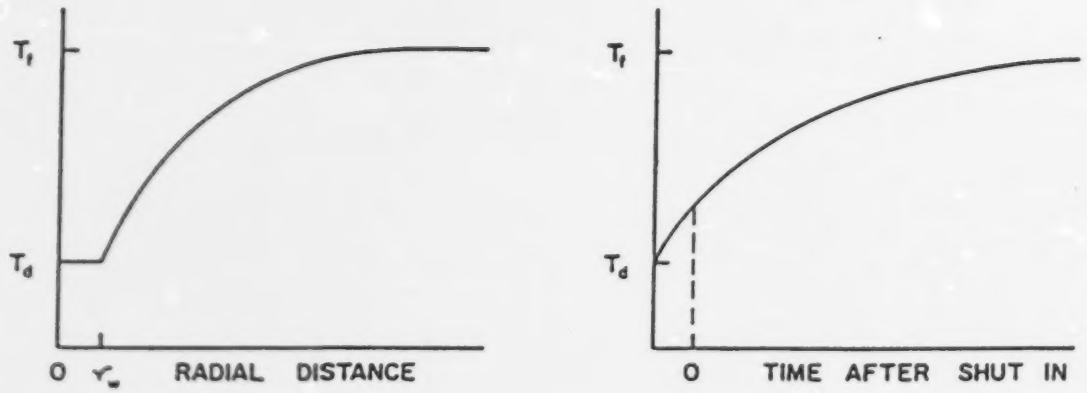



Figure 2.22 Schematic illustration of temperature profiles in the formation at the end of drilling and corresponding temperature variations in the borehole interval during testing for conditions of drill fluid temperature (a) greater than formation temperature and (b) less than formation temperature

period. These temperature variations result in significant pressure changes during pulse testing (i.e., under shut-in conditions), especially in low permeability formations, as a consequence of the thermal expansion (or contraction) of the fluid.

If a significant temperature change (i.e., 1 or 2°C) occurs over the testing period in response to a temperature differential between the rock and the borehole, the pressure in the interval will change due to the variation in temperature. The thermal expansion coefficient of water varies significantly over the potential range of subsurface temperatures. Consequently, the observed pressure change depends significantly on the temperature. Figure 2.23 illustrates the change in pressure due to a temperature change in a closed system (i.e., hypothetical interval with zero permeability). The graphs are calculated using the thermal expansion coefficient of water at the temperatures 20, 40, 60 and 80 °C. These graphs serve to illustrate the possible magnitudes of pressure change resulting from thermal expansion or contraction of the borehole fluid. For instance, at a temperature around 40°C, a 2° increase in temperature in a zero permeability shut-in interval will result in a pressure increase (expressed as hydraulic head) of about 180 metres. In an actual geologic test interval, the increase in pressure due to a temperature rise is moderated somewhat by the pressure dissipation into the formation.

The pressure response in a high permeability zone is relatively insensitive to temperature changes. The insensitivity is due to the relative ease with which a high permeability zone dissipates any pressure change, providing the zone has sufficient lateral extent to accommodate the pressure changes. It should be noted that testing of higher permeability zones would utilize slug tests which are not

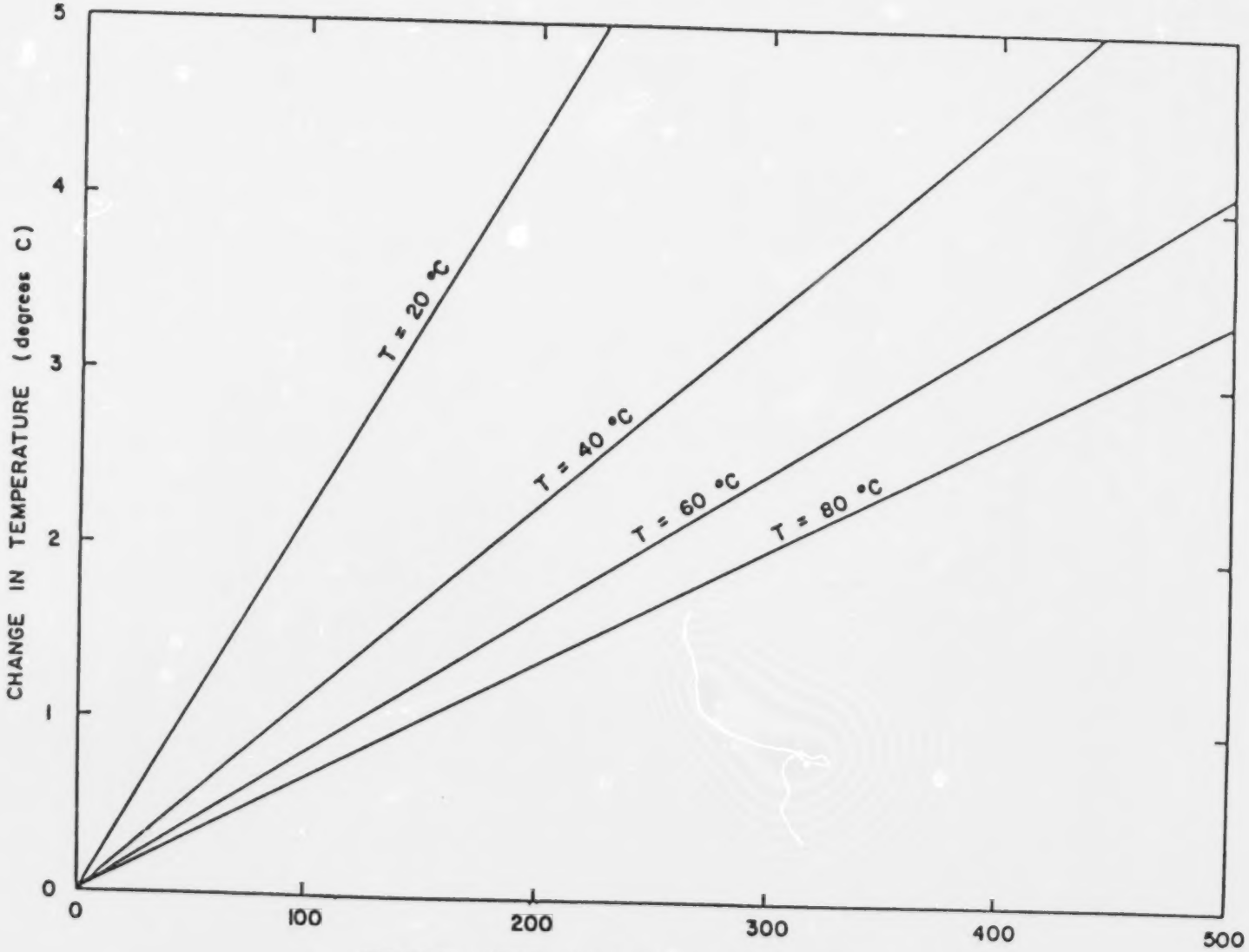



Figure 2.23 Induced hydraulic head change from a temperature change in the test interval at initial temperatures of 20, 40, 60 and 80 °C (assuming no fluid loss to the formation) From GTC, 1983

conducted under shut-in conditions, and therefore would not involve thermally-induced pressure effects.

For illustrative purposes, thermally-induced pressure effects on pulse tests are shown schematically in Figures 2.24 and 2.25 for the pressure conditions of formation pressure ($P_f$) greater than annulus pressure ($P_a$). It is important to note that the measured borehole pressure response curve under varying temperature conditions may have the same shape as a normal pressure response curve for tests of several hours to several days in length. However, analysis of such pressure response curves using type-curve analysis methods would yield incorrect permeability and formation pressure estimates, particularly in low permeability formations. The use of such analysis methods under varying-temperature conditions is incorrect since the basis of these methods ignores the important pressure and temperature relationships that must be included.

Pulse injection or withdrawal tests are conducted on lower permeability zones since they allow the tests to be conducted more rapidly as compared to slug injection or withdrawal tests. Varying-temperature conditions existing in any zone requiring pulse tests will likely result in significant thermally-induced pressure responses. In situations where a section of borehole has been open to annulus conditions for a long period of time and inflows or outflows over that section are minimal, a quasi thermal equilibrium will likely be reached between the formation and the borehole fluid at any particular depth. Under these conditions, thermally-induced pressure effects will be minimized.

2.1.3.3 Equipment Compliance

With any downhole packer/transducer equipment configuration, there is a certain amount of non-rigidity or compliance in the equipment. The equipment compliance can be related to a

PRESSURE RESPONSE DURING PULSE INJECTION TEST
IN LOW PERMEABILITY FORMATION

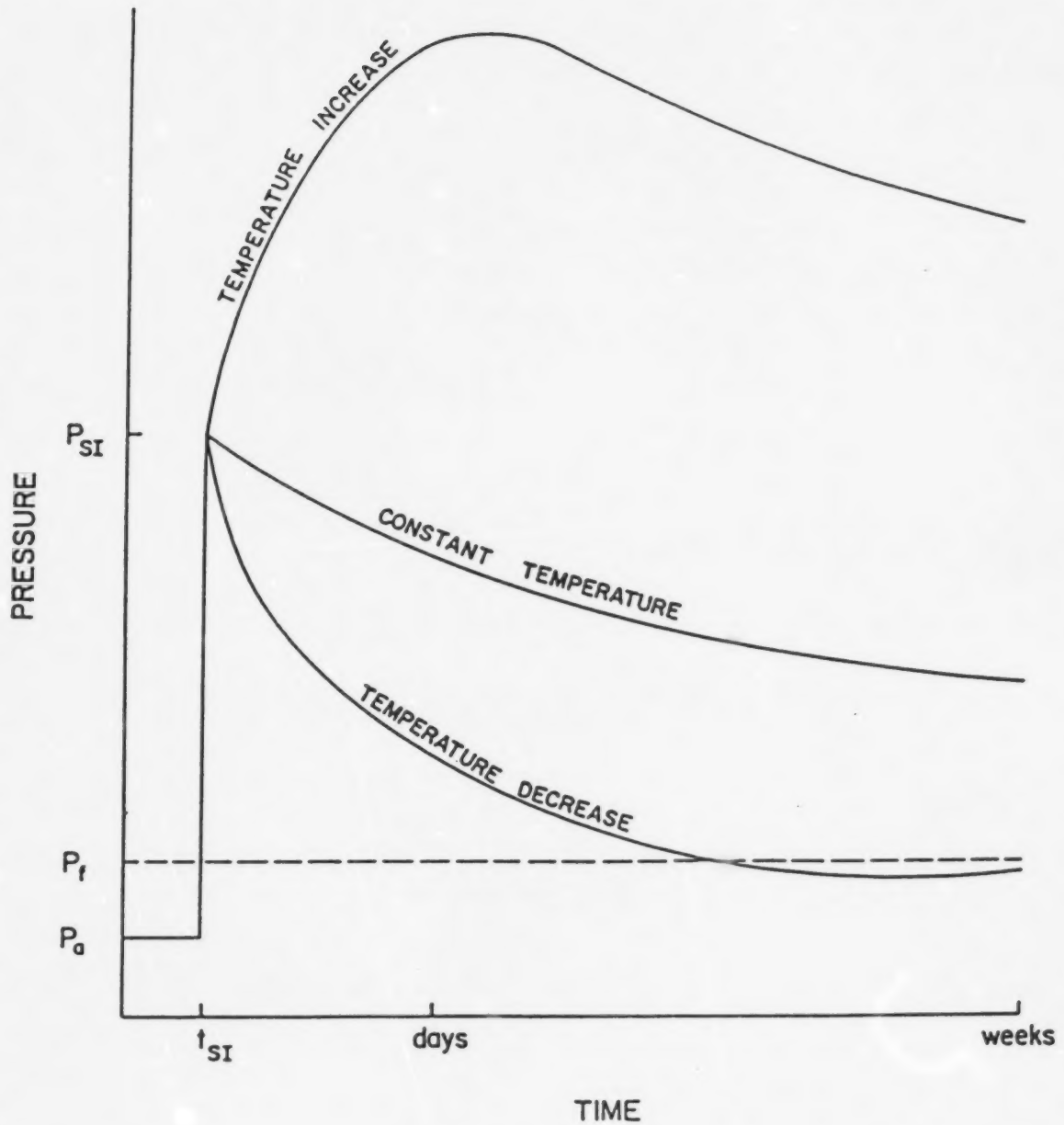



Figure 2.24 Schematic pressure responses during a pulse injection test in a low permeability formation for conditions corresponding to a temperature increase, a constant temperature and a temperature decrease

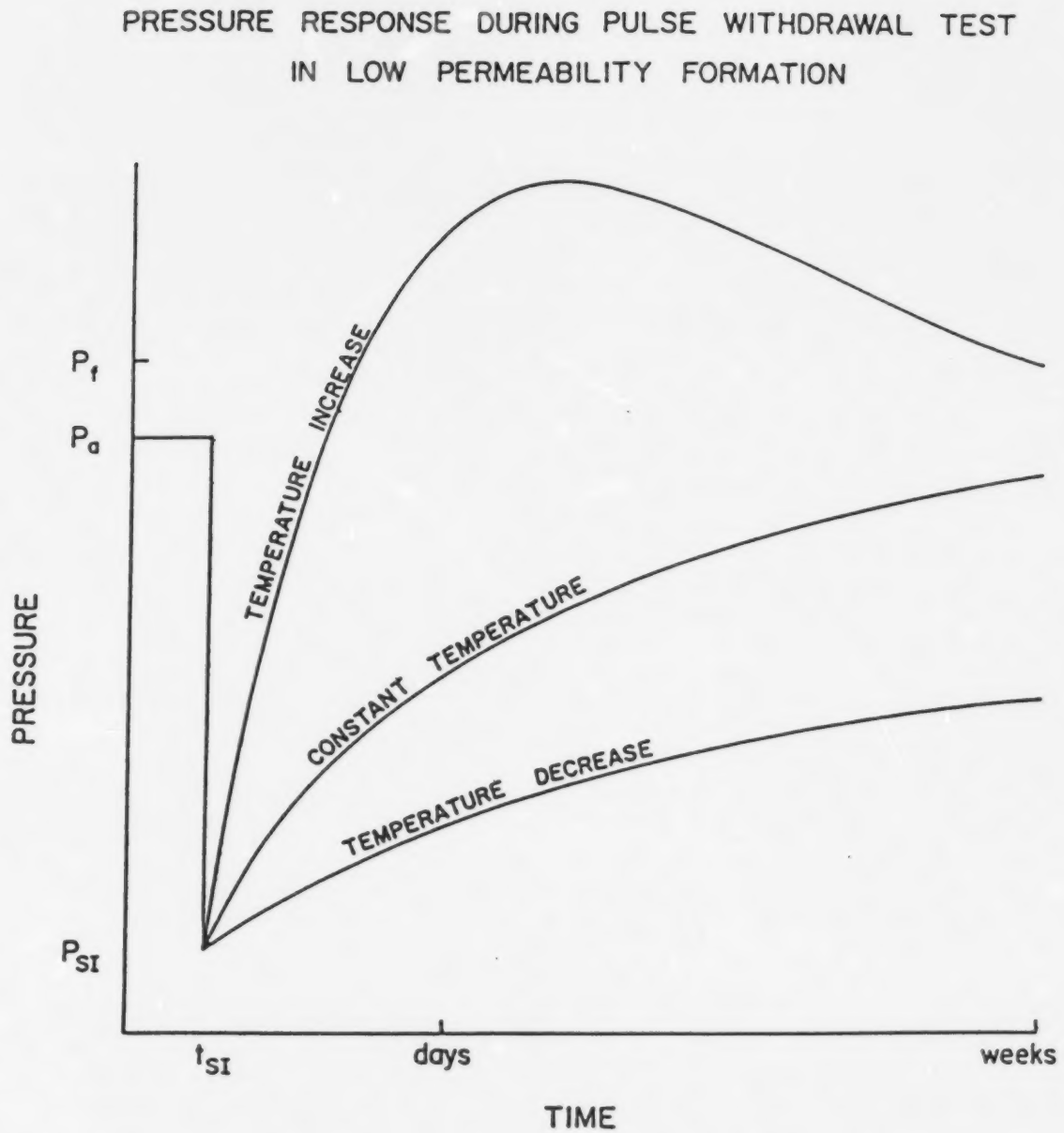



Figure 2.25 Schematic pressure responses during a pulse withdrawal test in a low permeability formation for conditions corresponding to a temperature increase, a constant temperature and a temperature decrease

number of things, including cold-flow of packer rubber, entrapped air in the equipment located in a test interval, machined tolerances on the seating and positioning of various steel components and O-ring or other seals, etc. For the case of a tubing string single packer configurations the schematic diagram in Figure 2.26 can be used to illustrate some of the possible equipment related effects. Subsequent to packer inflation, the "shut-in" movement, for instance, of the Lynes tool results in a "shut-in squeeze" which is a pressure increase in response to a small fluid injection into the test interval. In practice, the shut-in movement results in a weight of one to several tons transferred to the tubing string which closes the shut-in valve. In addition to the shut-in "squeeze" the added weight must, of course, be supported by the packer, which is inflated against the borehole wall. Since the packer and downhole system are not totally rigid, the increased weight tends to "compress" the isolated interval.

Examples of the increased pressure, which can result from weight transfer, and consequently a reduction in the isolated interval length are presented in Figure 2.27. It can be seen that if the length of a 10 metre test interval of a low permeability zone is reduced (or "squeezed") by 1 cm due to equipment compliance, the resulting increase in hydraulic head is more than 200 metres.

In order to develop a quantitative capability for analyzing pressure test data which include compliance effects, the magnitude and shape of the compliance function must be known. This requirement can be met by observations of pressure responses within impermeable lengths of casing or in "simulated boreholes" over a range of temperatures and applied loads. In the simplest case, such as the Lynes system, the compliance can be treated as a step function and either considered as part of the initial shut-in squeeze or

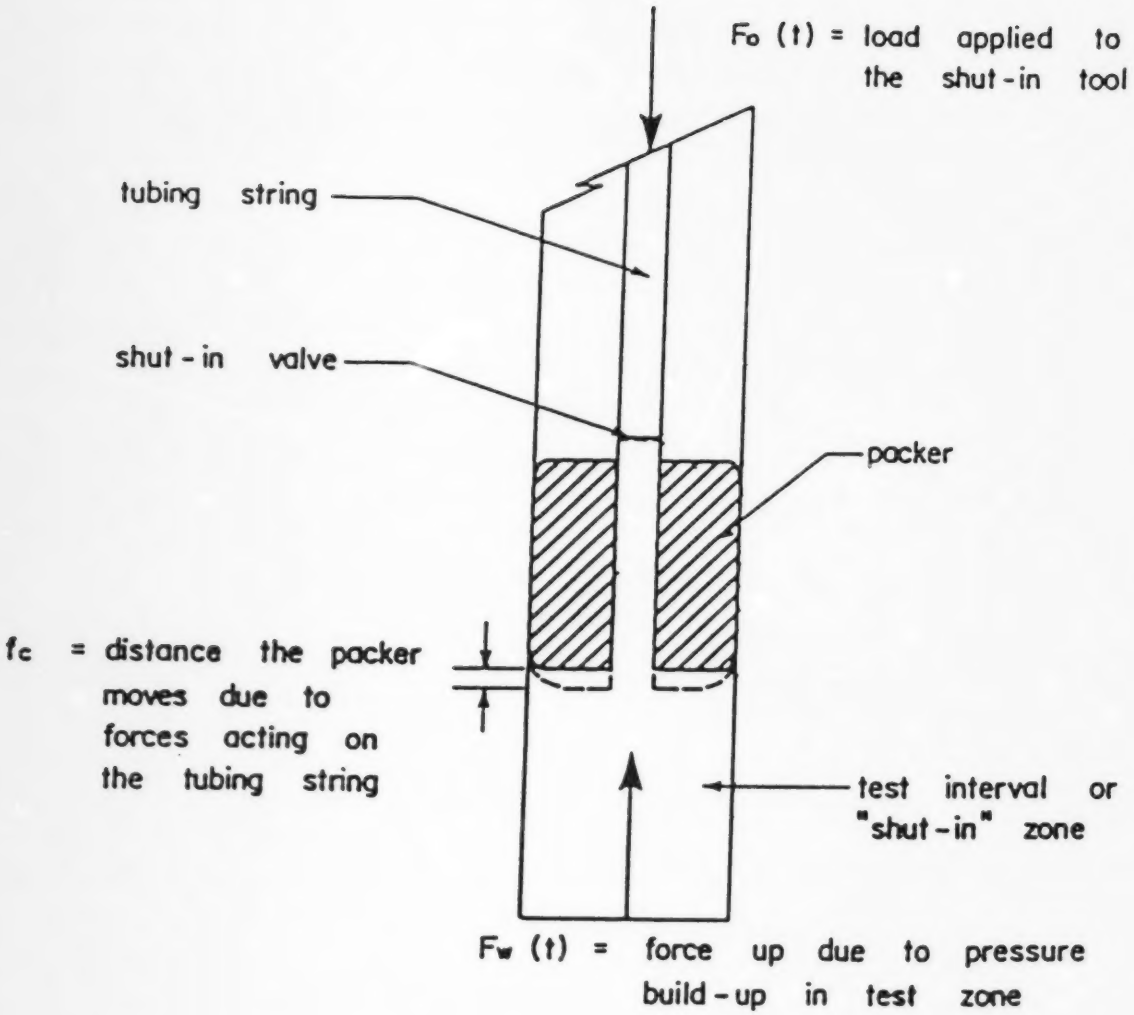

$F_o$ (t) = load applied to the shut-in tool
tubing string
shut-in valve
packer
$f_c$ = distance the packer moves due to forces acting on the tubing string
test interval or "shut-in" zone
$F_w$ (t) = force up due to pressure build-up in test zone

Figure 2.26 Schematic of applied forces in single packer test configuration (from GTC, 1983)

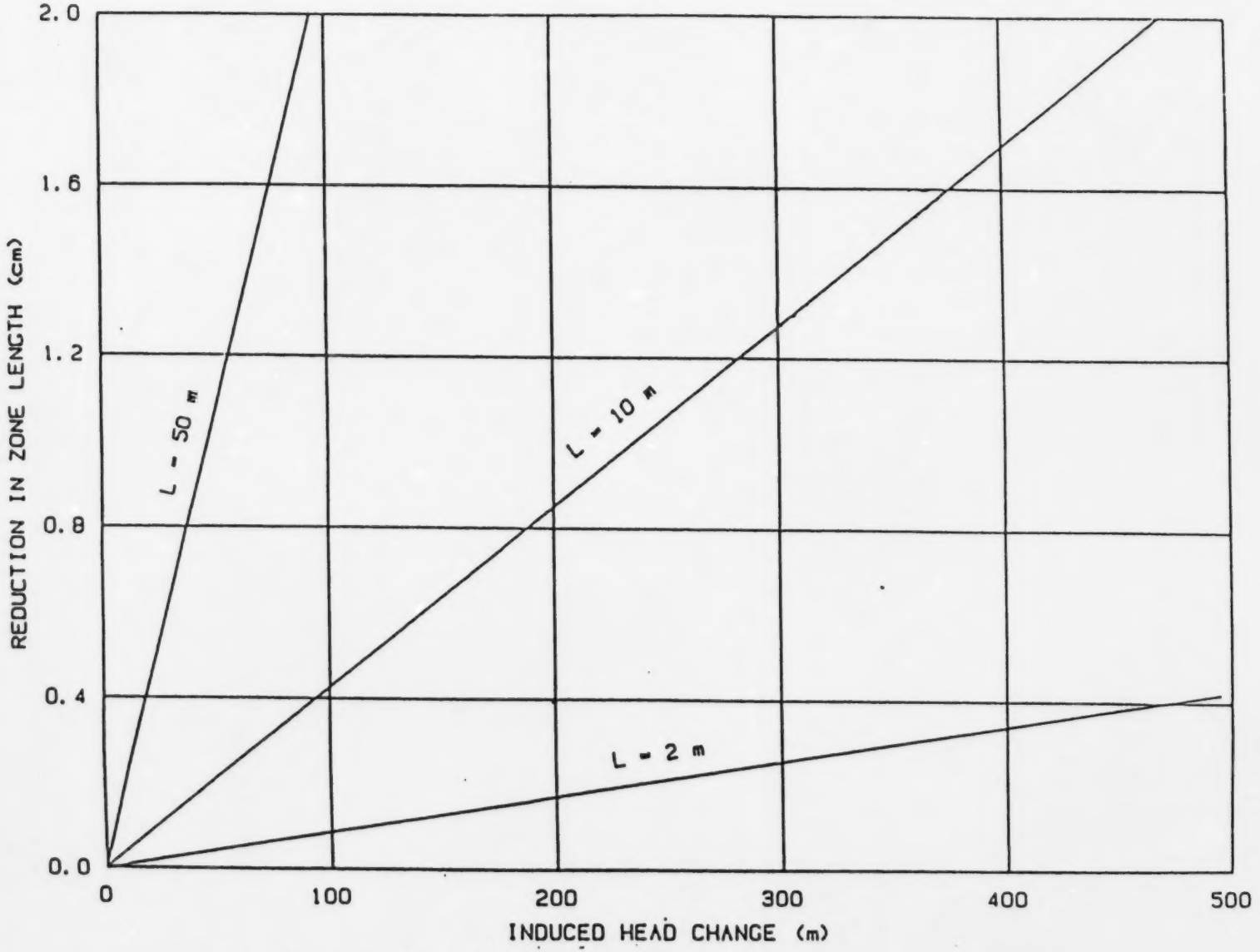



Figure 2.27 Induced hydraulic head increase from a reduction in test interval length for initial packed off intervals of 2, 10 and 50 m (assuming no fluid loss to the formation) From GTC, 1983

applied subsequently. In more complicated cases the compliance is regarded as a time, pressure and load dependent mathematical function, with the shape and magnitude defined by several compliance related parameters. Pressure responses observed during a long-term single packer (Lynes) test suggests that the bulk of the equipment compliance effects occur relatively rapidly (Figure 2.28).

### 2.1.4 Concluding Remarks

The preceding discussion has been presented for the purpose of demonstrating the difficulties associated with measurement of hydrogeologic parameters as well as illustrating the current research directions being undertaken. Low permeability measurements are, in particular, of interest in current research programs. At present the hydraulic information obtained can be presented in many ways. One way is the Hydrogeological Reconnaissance Log (H-Log) currently employed in the Canadian and Swiss radioactive waste disposal programs (Figure 2.29). The H-Log is used as the basis for hydrogeologic interpretations, in that the fundamental hydrogeologic information is presented in relatively uninterpreted form.

The reliability of permeability and hydraulic head information, is a matter of considerable importance. The measurements should be made under a strict quality assurance (QA) procedure to allow for consistency within and between data sets. The hydrogeologic data acquisition program employed in the Swiss program is an example of the type of QA procedures which can be employed during data acquisition phases of investigations. Guidelines for in-situ pressure measurement using the Lynes equipment have been developed specifically for the Swiss program. These guidelines and QA procedures have been reproduced in full in Appendix 3 to provide the reader with an understanding of the detailed evaluation which must go into a

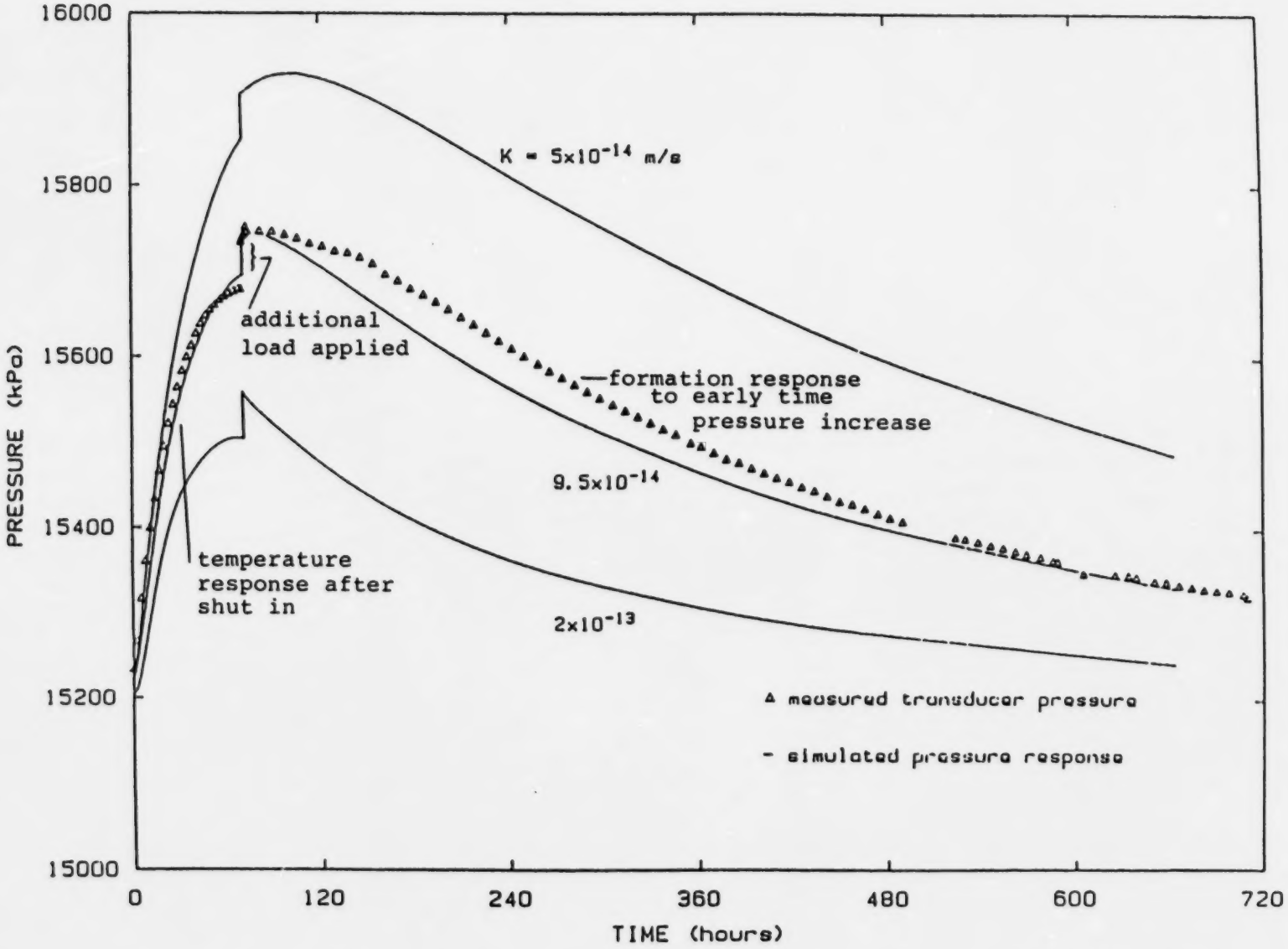



Figure 2.28 Measured and simulated pressure responses for a formation pressure of 15179 kPa with hydraulic conductivities of $5 \times 10^{-14}$ m/s, $9.5 \times 10^{-14}$ m/s and $2 \times 10^{-13}$ m/s (from GTC, 1984)

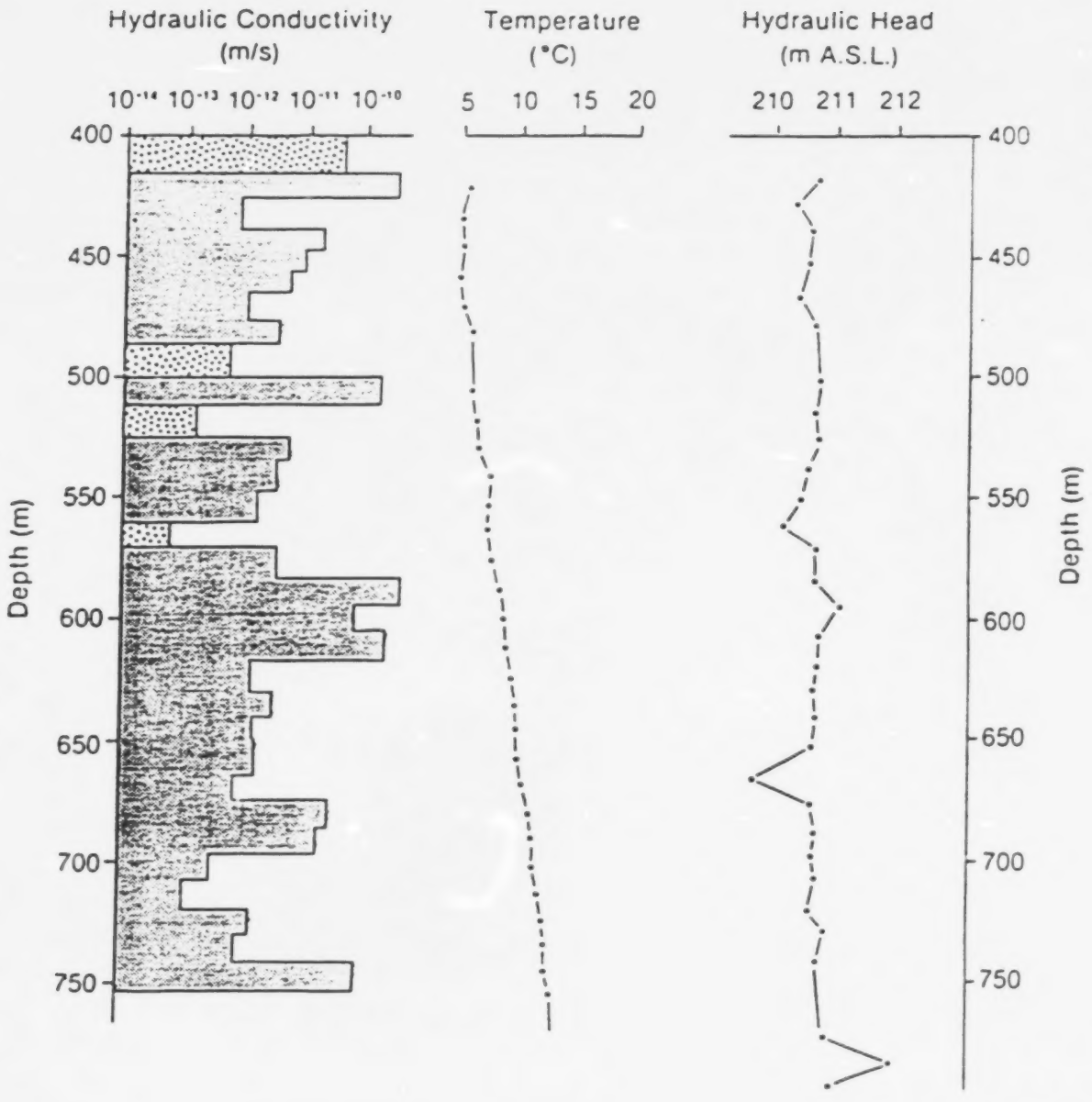



Figure 2.29 Schematic Diagram of Hydrogeologic Reconnaissance Log in Use in Canada and Switzerland (from GTC, 1982)

single pressure or hydraulic head measurement. It is correct to state that pressure and permeability measurement at depth in low permeability formations is a difficult and developing technology, still in the research stages.

## 2.2 Solute/Radionuclide Transport

Migration of radionuclides in any hydrogeologic system generally considers five basic processes.

1. Advection - radionuclides are transported with the ground water as it moves through a system of pores or fractures in response to the hydraulic gradient. The advective capacity of a hydrogeologic system is usually described by the average linear ground water velocity , which can be calculated from the Darcy velocity (Q/A) (eqn. 1.1) and the effective porosity of the hydrogeologic medium,

$$\frac{Q}{A} = K \frac{\Delta h}{\Delta L} \qquad (2.11)$$

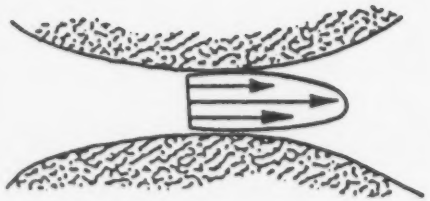
$$V = \frac{K \frac{\Delta h}{\Delta L}}{\theta} \qquad (2.12)$$

where $\theta$ is the effective porosity. The cross-sectional area of flow, A, must clearly be modified by the effective porosity, which is the amount of interconnected intergranular or fracture pore space that contributes to ground water flow.

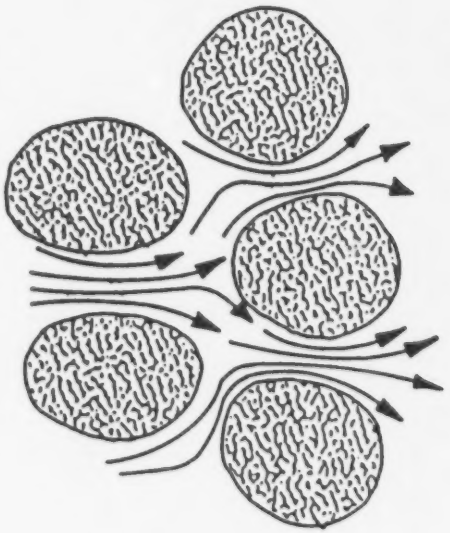
2. Mechanical Dispersion - the dispersive flux results from the variations of water velocity which occur across and within an individual pore or fracture opening as well as due to the interconnections and non-linearities of the porous or fracture system. This process is perceived on a macroscopic scale as an effective mixing.

   The microscopic nature of dispersion is illustrated in Figure 2.30.

MIXING IN INDIVIDUAL PORES



MIXING OF PORE CHANNELS



MIXING BY MOLECULAR DIFFUSION

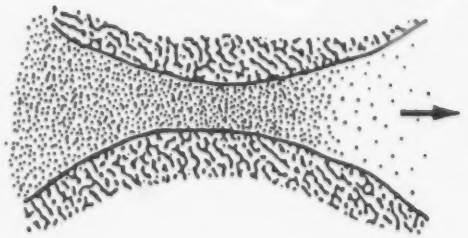


Figure 2.30 Dispersion Mechanisms in Porous Media (Pickens et al., 1978)

The dispersive flux essentially accounts for the mixing which occurs within an REV due to a) the variable velocity profile within an individual pore channel, b) mixing between pore channels and c) mixing by molecular diffusion in small or slow moving pore channels. Figure 2.31 extends the same concept to larger scales, eventually to the megascopic scale which could be represented by several different geologic units. It is clear that different velocities which occur in each unit can result in different arrival times at a sampling position, for instance a well. Figure 2.32 illustrates the effect of layers within an aquifer system which results in variable solute arrival times at a sampling position.

The conceptualization of dispersion in a fractured medium is strongly dependent on the scale to which the dispersion concept is applied. The dispersivity within a single fracture is likely small and dependent on the aperture, roughness, and type and frequency of bridging materials within the fracture plane. Considering only the mechanical component of dispersion, a fracture with enough porous gouge or bridging material will have velocity variations similar to those of a small permeable lens, bounded by virtually impermeable layers on either side. A fracture without bridging material and with completely smooth planar surfaces will have velocity variations parabolically distributed between the two planes. Recent laboratory studies (Los Alamos; Argonne National Laboratories) are known to be investigating the occurrence of "ribbon flow" or preferred channels within the plane of a single fracture. The consequence of this type of flow is also a variation in the velocity field, which is effectively dispersion.

Single fractures are generally discontinuous in the plane of the fracture, and hydraulic continuity within a rock mass is provided by interconnection with other fractures with

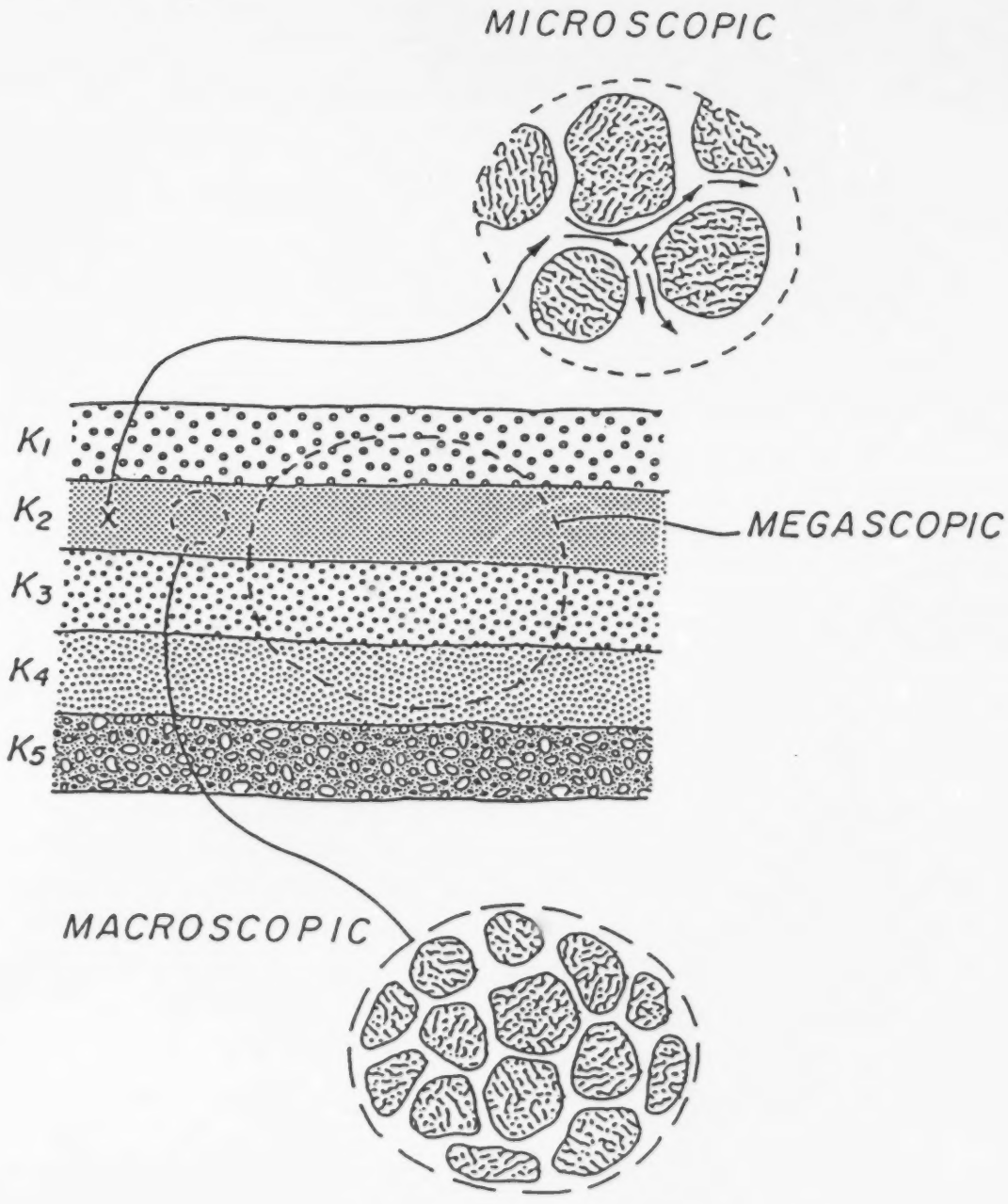



Figure 2.31 Scale Definitions in a Porous System (Pickens et al., 1978)

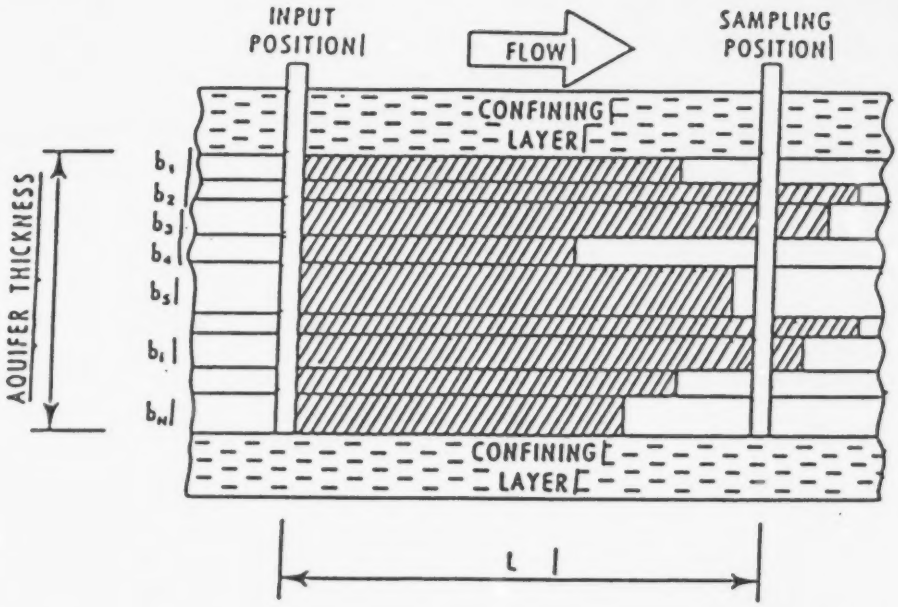



Figure 2.32 Effect of Layering or Zones of Different Hydraulic Properties on Solute Transport Arrival at a Sampling Position (from Pickens and Grisak, 1981)

nonparallel orientations. The dispersion analogy between porous and fractured media can be drawn by considering the fractures as permeable "lenses" (with individual characteristic dispersivities) within a less pervious medium. The dispersion that would be measured in any single- or multiple-well field tracer study (Webster et al., 1970; Grove and Beetem, 1971; Claassen and Cordes, 1975) is dependent on the number and interconnection of the permeable lenses and their individual characteristics. The resultant parameter of dispersivity, using conventional advection-dispersion analysis, will be that which conveniently provides the "characteristic mixing length" for the particular case or sampling interval and transport distance under consideration. Discussions of the complexity of the dispersion process in fractured media are further presented by Castillo et al. (1972) and Krizek et al. (1972).

The net effect of the dispersion process in porous or fractured media is to modify the solute arrival or breakthrough curves at a sampling position as illustrated in Figure 2.33.

3. Diffusion

As noted in 2 above, mixing occurs on the microscopic level due to the process of molecular diffusion. At the macroscopic or megascopic level in a porous medium system, diffusion of solutes from permeable aquifers into adjacent lower permeability media may also be a significant process. In a fractured medium with a porous matrix, molecular diffusion can take place both in the fluid-filled fractures and in water contained in the pores of the matrix. When solute diffuses into the matrix pores its movement along the fracture will be retarded with respect to the flux of water in the fracture since, in general, water flows more rapidly in fractures than through the matrix. Under some

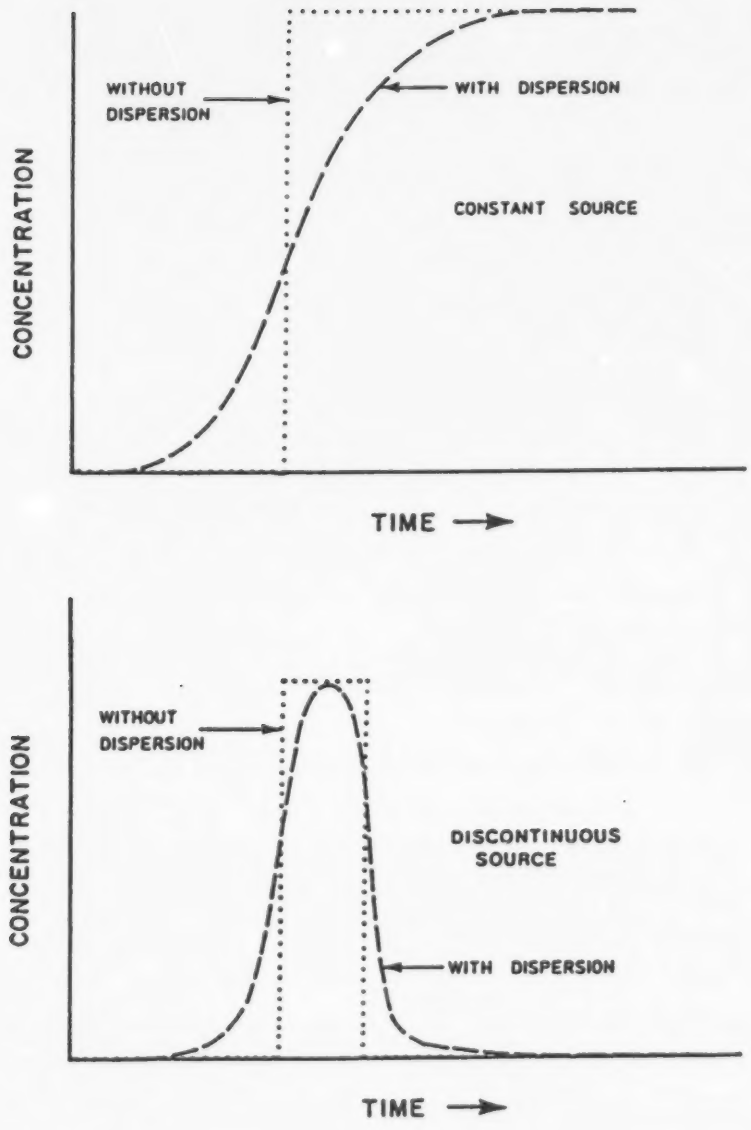


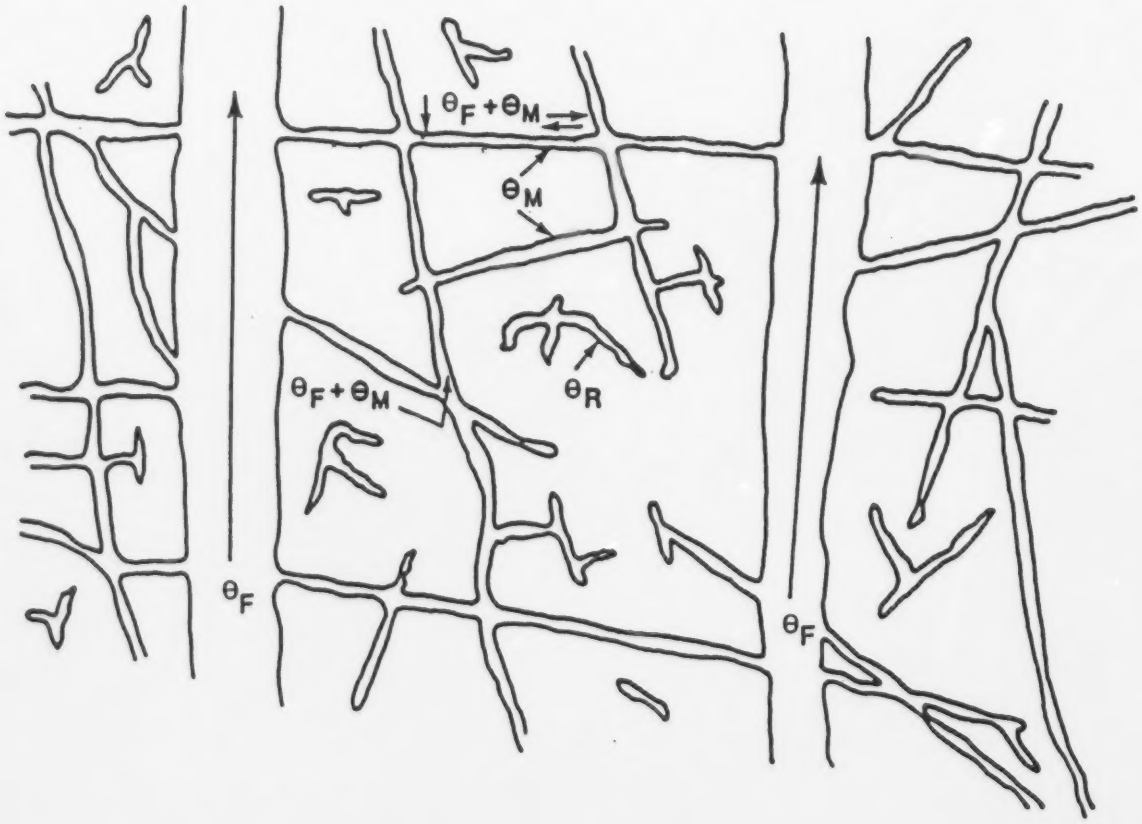

Figure 2.33 Solute Breakthrough Curves at a Sampling Point With and Without Dispersion

combinations of small fracture aperture and low hydraulic gradient along fractures axes, diffusion into the rock matrix may be the dominant process affecting solute transport. A discussion of the principles of diffusion in geologic media is presented in Appendix (2).

A schematic drawing of the various types of pore spaces present in a fractured crystalline rock (after Norton and Knapp, 1977) is presented in Figure 2.34. The flow porosity, $\theta_F$, is essentially the effective porosity as defined in 1. above. The flow porosity is the interconnected fracture system through which active ground water movement occurs. The matrix porosity, $\theta_m$, is the interconnected porosity of the rock matrix between the flow porosity fractures resulting from small cracks along grain boundaries and in individual crystals. The matrix porosity is the porosity available for the diffusion of solute from the flow porosity. The residual porosity, $\theta_R$, is the remaining unconnected pore space, such as individual isolated cracks or perhaps cooling-related features such as fluid inclusions.

In crystalline rocks, fractures are the principal paths of ground water flow and solute transport, although the matrix adjacent to the $\theta_F$ fractures plays an important part in the overall solute transport process. Solutes diffuse within the pore space available in the solid rock matrix. The process of solute diffusion from a $\theta_F$ fracture into the adjacent matrix is schematically illustrated in Figure 2.35 (GTC, 1983). The effect of matrix diffusion is to provide solute "storage" with the rate of change of storage within the matrix related to Fick's second law of diffusion.

At the single fracture scale, transport in a fractured medium can be considered to be dominated by advection in the fracture and moderated by matrix diffusion processes. At the




$\theta_F$ FLOW POROSITY
$\theta_M$ DIFFUSION POROSITY
$\theta_R$ RESIDUAL POROSITY

Figure 2.34 Types of Porosity in Fractured Media (after Norton and Knapp, 1977)

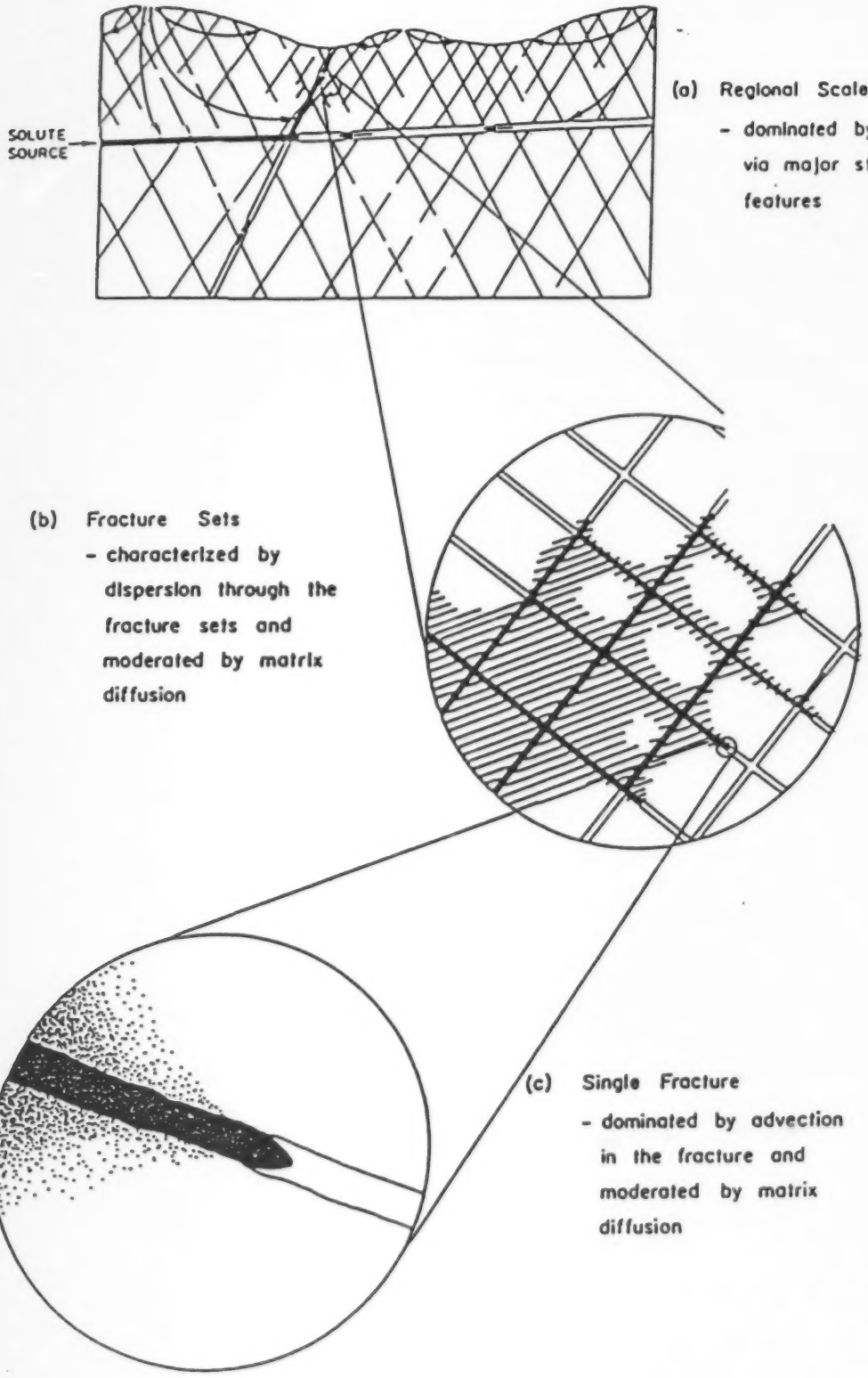

SOLUTE SOURCE
(a) Regional Scale
- dominated by advection via major structural features
(b) Fracture Sets
- characterized by dispersion through the fracture sets and moderated by matrix diffusion
(c) Single Fracture
- dominated by advection in the fracture and moderated by matrix diffusion

Figure 2.35 Solute Transport Mechanisms in Fractured Media (from GTC, 1983)

scale of multiple fractures or fracture sets, the solute transport processes are predominantly advection and dispersion through the fractures sets with some moderation by matrix diffusion. At the macroscopic or regional scale, transport through fractured media is likely dominated by advection in relatively major structural features.

The net effect of matrix diffusion is to retard the arrival of a solute, which is moving with the water in the $\theta_F$ fractures, at any point along the $\theta_F$ fracture. Minerals that coat the $\theta_F$ fractures, such as secondary oxides, carbonates, silicates etc., may alter the near-fracture diffusion process by creating a zone with reactive properties, porosities and diffusion coefficients different from those within the unaltered matrix. The diffusion process is, of course, reversible and if the source of solute to the $\theta_F$ fracture is discontinued the concentration gradient will reverse, causing solute to move from the matrix into the $\theta_F$ fracture.

Reactive matrix diffusion, or removal of the diffusing solute from the liquid by sorption onto the solid phase or chemical reaction, reduces the velocity of a solute in the $\theta_F$ fracture even more than it would be reduced by nonreactive diffusion. This phenomenon can be explained by considering the solute concentration gradient between the $\theta_F$ fracture and the first few $\theta_m$ matrix pores. As the reactive solute diffuses into the matrix pores, some of the solute is removed from solution by adsorption on the particle surfaces in the matrix. If diffusion is considered to be occurring within the solution phase only (i.e., the effect of adsorbed ions is neglected) then removal of ions from solution maintains a high concentration gradient and a strong solute flux from the $\theta_F$ fracture into the rock matrix. The net result of the

continued strong flux of solute into the matrix is to reduce the average solute velocity in the fracture.

4. Geochemical Retardation - transfer across phase boundaries due to adsorption, precipitation and ion exchange will retard advective and dispersive solute transport. The retardation of a reactive solute is usually considered to occur as illustrated in Figure 2.36 (from Pickens et al., 1981).

   The difference between the "predicted" and "measured" reactive transport curves is usually interpreted as an increased spreading due to either physical nonequilibrium effects (such as dead end pore space diffusion) or to the kinetics of geochemical reactions.

5. Radioactive Decay

   Radionuclides decay to daughter isotopes which may be either radioactive or stable end products. The process of radioactive decay serves in some instances to limit migration of radioactive solutes, while in others, for instance where the daughter products are long-lived radioisotopes, radioactive decay can result in increased retention of radionuclides in a ground water system.

   In the case of fracture transport, when matrix diffusion of a radionuclide occurs, it may be possible to develop a "steady-state" profile for the radioactive element if the source remains constant. This profile is a consequence of the decay and apparent "loss" of the radionuclide as it diffuses away from the fracture, which is balanced by the input of nondecayed radionuclides at the fracture/matrix interface. Steady-state profiles have been calculated for short-lived radionuclides (Tang et al., 1981) under relatively simple physical and chemical boundary conditions.

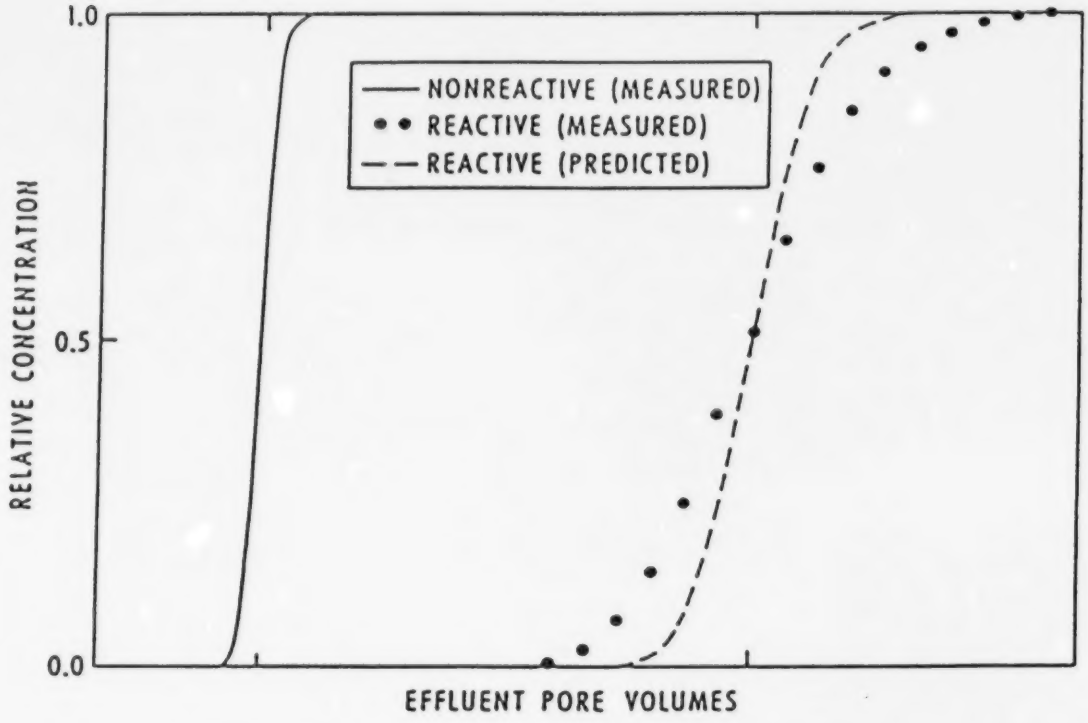



Figure 2.36 Schematic Breakthrough Curves for Reactive and Nonreactive Solute Transport (from Pickens et al., 1981)

The longer-lived nuclides would likely be more subject to variability in the natural system (i.e., adjacent fractures, changing flow regimes) and would be less likely to develop steady-state profiles.

### 2.2.1 Governing Equations

The solute transport equation describing the transient concentration distribution of a dissolved constituent being transported in a saturated medium can be written in two dimensions as (Pickens and Grisak, 1979):

$$L(C) = \left(\theta + \rho_b K_d\right) \frac{\partial C}{\partial t} - \frac{\partial}{\partial x}\left[\theta D_{xx} \frac{\partial C}{\partial x} + \theta D_{xz} \frac{\partial C}{\partial z}\right] - \frac{\partial}{\partial z}\left[\theta D_{zx} \frac{\partial C}{\partial x} + \theta D_{zz} \frac{\partial C}{\partial z}\right] + q_x \frac{\partial C}{\partial x} + q_z \frac{\partial C}{\partial z} + \lambda \left(\theta + \rho_b K_d\right) C + Q^C = 0 \qquad (2.13)$$

where $\rho_b$ is the bulk density of the medium, $K_d$ is the distribution coefficient, C is the concentration of the constituent in solution, $D_{xx}$, $D_{xz}$, $D_{zx}$ and $D_{zz}$ are the components of the hydrodynamic solute dispersion tensor, $\lambda$ is a first-order reaction constant and $Q^C$ is the mass rate per unit volume for addition or removal at a source or sink. This form of the equation includes the effects of dispersion, advection, reversible instantaneous sorption described by a distribution coefficient, a first-order reaction such as radioactive decay, and a source or sink for the dissolved constituent. It is also possible to represent nonlinear sorption relationships instead of the linear function given by the distribution coefficient.

The components of the hydrodynamic solute dispersion tensor include the effects of mechanical dispersion and molecular diffusion and are written (adapted from Reddell and Sunada, 1970):

$$\theta D_{xx} = \frac{\alpha_L q_x^2}{\bar{q}} + \frac{\alpha_T q_z^2}{\bar{q}} + \theta D^* \qquad (2.14a)$$

$$\theta D_{zz} = \frac{\alpha_T q_x^2}{\bar{q}} + \frac{\alpha_L q_z^2}{\bar{q}} + \theta D^* \qquad (2.14b)$$

$$\theta D_{xz} = \theta D_{zx} = \left(\alpha_L - \alpha_T\right) \frac{q_x q_z}{\bar{q}} \qquad (2.14c)$$

where $\alpha_L$ and $\alpha_T$ are the longitudinal and transverse dispersivities, respectively, of the medium, $\bar{q}$ is the resultant Darcy velocity, and $D^*$ is the diffusion coefficent of the solute in the porous medium. This type of formulation for the dispersion coefficients in the solute advection-dispersion equation is in widespread use in numerical modeling studies in fractured and porous media (Bredehoeft and Pinder, 1973; Pinder, 1973; Konikow, 1976, Pickens and Lennox, 1976). The dispersivities $\alpha_L$ and $\alpha_T$ are usually assumed to be identical for the mechanical dispersion of both heat and solutes although it is conceivable that the characteristic mixing length represented by the value of dispersivity may be different for the heat and solute transport cases.

The initial conditions necessary for the solution of Equation (2.13) are:

$$C(x,z,0) = \tilde{C}_o\ (x,z) \qquad (2.15)$$

where $\tilde{C}_o$ are initial prescribed concentration. The boundary

conditions can be Dirichlet (prescribed concentration) expressed as:

$$C(x,z,t) = \tilde{C}\ (x,z,t) \qquad \text{on } \Gamma_3 \qquad (2.16a)$$

Neumann expressed as:

$$\left[\left(\theta D_{xx} + \theta D_{zx}\right) \frac{\partial C}{\partial x} + \left(\theta D_{xz} + \theta D_{zz}\right) \frac{\partial C}{\partial z}\right] \bar{n} = 0 \qquad \text{on } \Gamma_4 \qquad (2.16b)$$

and Cauchy or mixed-boundary conditions expressed as:

$$\left[\left(\theta D_{xx} + \theta D_{zx}\right) \frac{\partial C}{\partial x} + \left(\theta D_{xz} + \theta D_{zz}\right) \frac{\partial C}{\partial z} + q_x C + q_z C\right] \bar{n} = \left(q_x \hat{C} + q_z \hat{C}\right) \bar{n} \qquad \text{on } \Gamma_5 \qquad (2.16c)$$

where $\tilde{C}$ are prescribed concentrations, $\Gamma_3 + \Gamma_4 + \Gamma_5 = \Gamma$, the total boundary of the system, $\hat{C}$ are prescribed concentrations of the solute at a solute influx boundary, and $\bar{n}$ is the outward normal vector to the boundaries $\Gamma_4$ and $\Gamma_5$.

Specifically for fractured media transport, the following simplified equation for a unidirectional flow field describes the principal radioactive solute transport process in fractured media.

$$L(C) = \left(\theta + \rho_b K_d\right) \frac{\partial C}{\partial t} - \frac{\partial}{\partial x}\left(\theta D_{xx} \frac{\partial C}{\partial x}\right) + q_x \frac{\partial C}{\partial x} - \frac{\partial}{\partial y}\left(\theta D_{yy} \frac{\partial C}{\partial y}\right) + \lambda\left(\theta + \rho_b K_d\right) C = 0 \qquad (2.17)$$

where x is the direction of flow and y is normal to the direction of flow. The radioactive decay term, $\lambda$, is a constant and $\theta$, $\rho_b$, $K_d$, $D_{xx}$ and $D_{yy}$ are spatially-varying parameters. Conceptually, the general equation 2.17 can be

applied at virtually all scales, implying however that the physical meaning of the terms at various scales is defined. For instance, at the scale of a single fracture, the second and third terms on the right-hand side of equation (2.17) describe classical dispersion and advection in the direction of flow, while the fourth term can be utilized to describe a diffusive flux of solute in or out of the matrix.

The dispersion coefficients can be simplified to

$$D_{xx} = \alpha_L V_x + D^* \qquad (2.18a)$$

$$D_{yy} = \alpha_T V_y + D^* \qquad (2.18b)$$

where $\alpha_L$ and $\alpha_T$ are the longitudinal and transverse dispersivities of the medium. The dispersion coefficient normal to the direction of flow (i.e., in the matrix) can be reduced to the effective molecular diffusion coefficient, $D^*$, by assuming the advective velocity in the matrix to be extremely small (i.e., negligible) compared to that in the fracture. It is also possible, however, to incorporate velocities in the matrix as well as dispersion terms by employing numerical solutions. By numerically solving equation 2.17 and discretizing both the fracture and the matrix, virtually all of the important solute transport processes in fractured media can be incorporated (e.g., Grisak and Pickens, 1980).

Although the physical and chemical processes can be translated into governing equations, considerable care must be exercised concerning the scale at which these equations are employed. For instance, the effective porosity for fractured crystalline rock at the regional scale is considerably less than the effective porosity for porous media. Therefore, the assumptions inherent in the governing equation concerning the amount of rock available for solute interaction or sorption may be erroneous when applied to fractured media (however, it is possible to show

that, at the extreme which considers the entire flow system domain as a Representative Elementary Volume, the assumption with respect to REV's is correct). For these reasons the parameters must be considered individually within the context of the scale of the conceptualization.

For practical purposes, the following list is considered fundamental as input data to allow the numerical or analytical application of the governing equations in a predictive quantitative assessment. A discussion of these parameters as they relate to various scales of investigation is presented in Grisak et al. (1985).

### 2.2.2 Parameter Evaluation (Data Required to Quantitatively Evaluate Radionuclide Transport)

#### 2.2.2.1 Ground Water Velocity

This parameter is either measured directly, perhaps by inference from environmental isotopes or other tracer studies, or calculated from the flow regime and the effective porosity (equations 2.11 and 2.12). Tracer tests suitable for determining ground water velocity are natural gradient tracer tests and borehole dilution tests. The natural gradient test involves introducing a tracer at one or more wells and monitoring the tracer arrival at points or wells downgradient in the flow system (Figure 2.37). In addition, this type of test allows the determination of flow paths.

The borehole dilution method involves introducing a tracer in an isolated interval of a borehole or well and monitoring the change in tracer concentration with time as the tracer migrates into the formation under the influence of the advective transport of the ground water (Figure 2.38). The

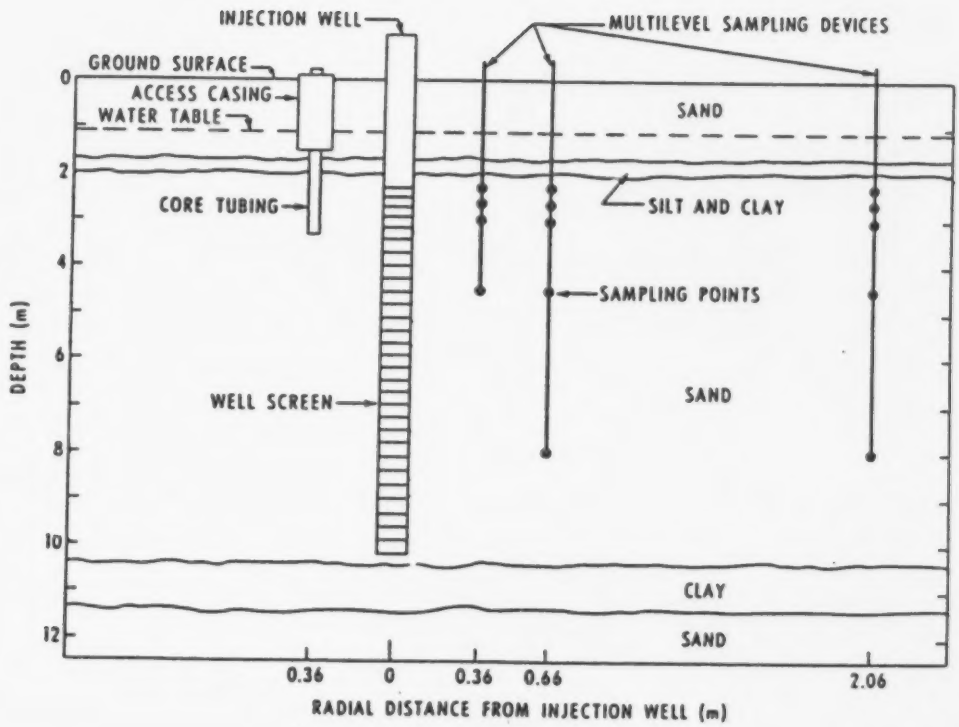



Figure 2.37 Schematic Diagram of Radial Injection/Withdrawal Test at CRNL Using Reactive and Nonreactive Tracers (from Pickens et al., 1981)

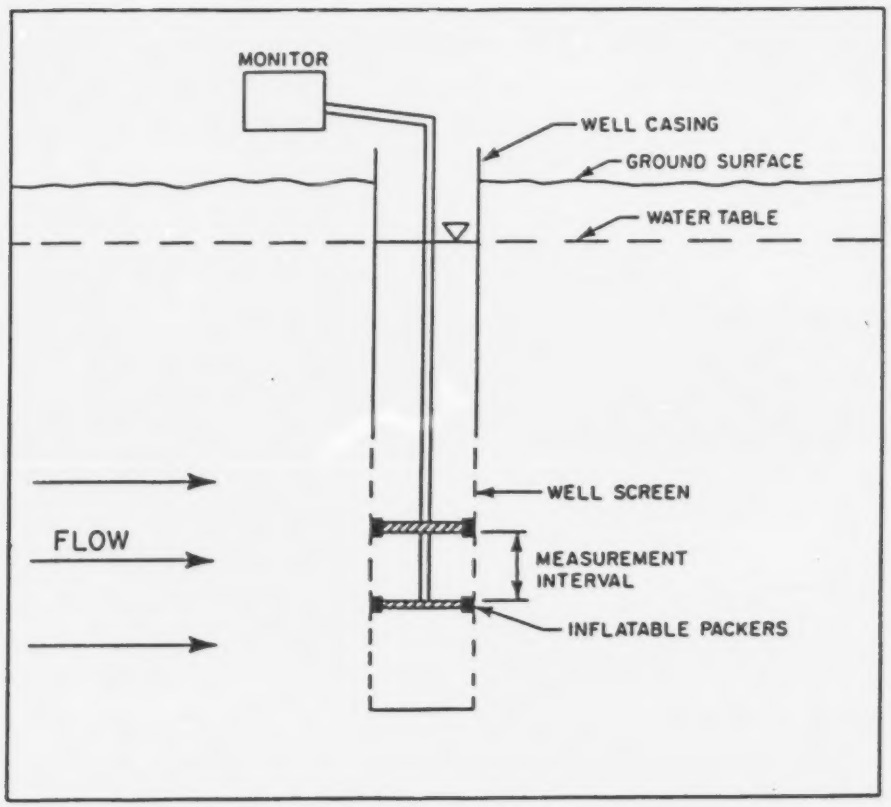



Figure 2.38 Schematic Diagram of Field Set-up for Borehole Dilution Test

effects of molecular diffusion must be considered for low ground water velocity situations (eg., less than several centimetres/day). Neglecting the effects of diffusion, the rate of tracer dilution in the packer-isolated interval of the borehole or well is a function of the ground water velocity through the expression

$$V_{GW} = \frac{V}{\theta A t} \ln \frac{C}{C_o} \qquad (2.19)$$

where $V_{GW}$ = ground water velocity in the formation
$V$ = volume of the borehole interval
$t$ = time
$C$ = concentration at time t
$C_o$ = concentration at time zero
$\theta$ = porosity

Usually a correction factor is applied to this seepage velocity in estimating the formation seepage (Darcy) velocity in order to compensate for the increased flow (streamline distortion) resulting from the presence of the borehole. For borehole dilution tests in porous media, the above velocity expression overestimates the formation seepage velocity by a factor of 2 or 3 (Drost et al., 1968). When natural gradient and borehole dilution tests are used in the same formation, an in-situ effective porosity of the formation can be determined.

For fractured rock, an expression for velocity measured by borehole dilution is (after Maini, 1971)

$$V = \frac{\pi d_w L_w}{8t \sum_i (2b)_i} \ln \frac{C}{C_o} \qquad (2.20)$$

where $d_w$ = borehole diameter

$L_w$ = packed off borehole interval

$\sum_i (2b)_i$ = total fracture aperture of i fractures

Tracer tests usually measure velocities that are representative of a small portion of the hydrogeologic system. This is the result of the large time and cost commitments to conduct large-scale tracer tests. Environmental isotopes have been used in some cases to determine regional scale migration rates or ground water residence times. The radioisotopes $^{14}C$, $^{3}H$ and the ratio of the uranium isotopes $^{234}U/^{238}U$ are the most common examples. One very important aspect of the use of environmental isotopes at the regional scale is in conjunction with the calibration of mathematical models of large-scale hydrogeologic systems. These calibrations are, in some cases, for the same time scale of interest to the prediction of the transport of radionuclides from a nuclear repository. Pearson, et al. (1983) have presented the results of such a study where a transport model was used to predict the behaviour of environmental $^{14}C$ in a regional hydrologic system.

The most widely used approach to estimating ground water velocities is by calculation from the measured and conceptualized distribution of permeability, hydraulic head and porosity within a hydrogeologic system. When the data obtained from a physical hydrogeologic investigation is assimilated using a two- or three-dimensional mathematical flow model, one of the model outputs is generally the spatial distribution of either the Darcy or ground water velocities.

Ground water velocity measurements in fractured media are not quite as straightforward as equations (2.11) and (2.12) would suggest. The scale of measurement is critical in fractured media. Within an individual fracture, the one-dimensional

flow velocity (in the limiting case of smooth planarity) may be considered to be parabolically distributed across the fracture, with the average velocity defined by

$$\bar{V} = \frac{\rho g\ (2b)^2 (dh/dx)}{12\mu} \qquad (2.21)$$

where $\bar{V}$ = average fluid velocity
$\rho$ = fluid density
$g$ = gravitational constant
$2b$ = fracture aperture
$dh/dx$ = hydraulic gradient
$\mu$ = dynamic viscosity of fluid

In a system of parallel, equally-spaced similar fractures, the average flow velocity is defined by

$$\bar{V} = \frac{\rho g (2b)^3 (dh/dx)}{12\mu\Delta} \qquad (2.22)$$

where $\Delta$ is the fracture spacing.

In a laboratory study of a fractured till block with two orthogonal sets of discontinuous fractures, Grisak et al. (1980) approximated the system for modeling purposes by developing a "representative" or "equivalent" fracture aperture and water velocity. This was accomplished by employing the concept of an "effective" fracture spacing which was calculated from observed flow rates through the column. The effective fracture spacing was calculated with an adapted form of the area-discharge relationships developed by Snow (1969) for parallel equally-spaced fractures. The effective fracture spacing $\Delta$ is written:

$$\Delta = \frac{\rho g\ (2b)^3 (dh/dx)\ A_T}{12\mu Q_T} \qquad (2.23)$$

where $Q_T$ = total volumetric flow rate
$A_T$ = total cross sectional flow area, including fractures and matrix

The above provided an extension from one fracture to the larger scale of sets of fractures, in the similar manner to which individual pore velocities in granular media are extended to the "average linear pore-water velocity" by the utilization of the Representative Elementary Volume (REV) concept. The conceptualization and scale extension proved adequate to model and evaluate the observed laboratory tracer test results for a fractured till sample (Grisak et al. 1980).

It can also be shown that, for a system with two fracture sets at equal spacing, the average flow velocity can be calculated using the Darcy analogy

$$\bar{V} = \frac{K\ (dh/dx)}{\theta} \qquad (2.24)$$

where $K$ = hydraulic conductivity of the bulk rock (or $K_{erm}$)
$\theta$ = effective porosity
and $\theta$ is defined by

$$\theta = \frac{2(2b)}{\Delta} \qquad (2.25)$$

Without proceeding through a rigorous development, it is clear that, at a scale which includes a sufficient number of fractures to encompass an REV of the flow regime, fractured media can be conceptualized for velocity terms using the equivalent porous media analogies and equations.

2.2.2.2 Dispersion Coefficient

As discussed in the introductory portions of this section, the physical meaning of the dispersion coefficient is highly dependent on the scale of conceptualization. In general terms, advection describes the mean transport rate of a solute, and hydrodynamic dispersion (molecular diffusion plus mechanical dispersion) explains the variation about the mean. If one could describe the three-dimensional flow pattern at the microscopic scale within a geologic medium, then it is conceivable that the transport of the solute could be described by the processes of advection and molecular diffusion alone. Since it is impossible to define the flow system at this scale, it is necessary to average the ground water velocity over some representative volume. The averaging, or the scale at which one chooses to define the flow system, can result in a scale dependence of the mechanical dispersion coefficient or the dispersivity. The parameter dispersivity, which is a measure of the dispersive properties of a hydrogeologic system, has traditionally been considered a characteristic single-valued property of the entire medium (Bear, 1972). In recent years, there have been field tracer studies which suggest that dispersivity is not a constant but rather depends on the mean solute travel distance or travel time and/or the scale of the system. Further, the nontransferability of dispersivities determined from laboratory tracer tests to field conditions has become well known. This nontransferability can be considered to be a consequence of change of scale and increased heterogeneity of the field in comparison to laboratory conditions.

There is a wide range of values for dispersivities reported in the literature obtained from the analysis of tracer tests or from simulation of contaminant transport cases. In general,

it can be considered that the magnitude of the dispersivity obtained is directly related to both the scale of definition of the flow system and the degree of heterogeneity of the geologic system at that scale. Conducting a small-scale tracer test to obtain a dispersivity for regional scale predictive purposes is not a constructive approach. Estimating a dispersivity by taking into account the heterogeneity of the geologic system and the spatial and temporal scales of the predictive requirements is considered to be a more useful approach. A summary of tracer tests in granular media from which dispersion coefficients (or dispersivity) have been obtained is presented in Table 2.2 (from Pickens and Grisak, 1981).

A review of tracer tests in crystalline rock, some of which have yielded dispersivity estimates, is presented in Table 2.3 (adapted from GTC, 1983).

#### 2.2.2.3 Matrix Diffusion Coefficient

The matrix diffusion coefficient can be measured in the laboratory (e.g., Walter, 1982) or in field studies (e.g., Birgersson and Neretnieks, 1982).

Laboratory and field tests can also be employed to evaluate many of the physicochemical parameters which directly affect the diffusion of solutes in fractured rocks. The most significant parameters (see Appendix 2) to be measured are (1) the total volume of interconnected pores with nominal diameters greater than approximately 0.1 m [i.e. the sum of the "diffusion porosity" and the "flow porosity" of the rock (Norton and Knapp, 1977)] (2) the effective (matrix) diffusion coefficients for the solutes of interest; and (3) the tortuosity, $\tau$, and constrictivity, c, of the interconnected pores (or, alternatively, the ratio of these parameters,

TABLE 2.2 Field Dispersivity Values in Granular Materials Obtained From Results of Tracer Tests

| Reference | Type of Test | Distance Between Wells, m | Slotted Length of Wells, m | Longitudinal Dispersvity, m |
|---|---|---|---|---|
| Theis (1963) (cited by Cole (1972) | natural gradient | 3500<br>4000 | n.r.<br>n.r. | 6<br>460 |
| Mercado (1966) | single-well injection-withdrawal<br>injection-withdrawal<br>observation well | n.a.<br>115 | 34<br>12-39 | 0.09-0.15<br>0.50-1.50 |
| Percious (1969) | single-well injection-withdrawal | n.a. | 30.5 | 0.08-0.25 |
| Wilson (1971) (cited by Robson (1974) | two-well nonrecirculating withdrawal-injection | 79.2 | 14.6 | 15.2 |
| Wilson (1971)<br>Fried et al. (1972) | single-well injection-withdrawal<br>single-well pulse (individual layers) | n.a.<br>n.a. | 15.2<br>0.25<br>layers | 0.25-0.33<br>0.1-0.6 |
| Kreft et al. (1974) | two-well pulse | 5-6 | n.r. | D/V = 0.18 |
| Robson (1974) | two-well recirculating withdrawal-injection | 6.4 | 27.4 | 15.2 |
| Fried (1975) | natural gradient<br>radial injection | 12<br>n.r. | n.r.<br>n.r. | 4.25<br>11.0 |
| G.E. Grisak (unpublished data, 1977) | single-well injection-withdrawal | n.a. | 1.0 | 0.29 |
| Rousselot (1977) | natural gradient | 60 | n.r. | 1 |
| Peaudecerf and Sauty (1978) | uniform gradient | 32.5 | n.r. | 1-2.7 |
| Sauty (1978) | two-well pulse | 9<br>6 | n.r.<br>n.r. | 6.9<br>0.3 |
| Sudicky and Cherry (1979) | natural gradient | 14 | point sampling | 0.01-0.22 |
| Sauty et al. (1979) | single-well injection-withdrawal<br>central well<br>observation wells | n.a.<br>13 | n.r.<br>n.r. | 1<br>0.18 x radius |
| Lee et al. (1980) | natural gradient | 6 | point sampling | 0.012 |
| Pickens and Grisak (1981) | single-well injection-withdrawal<br>central well<br>multilevel samples | n.a. | 8.0 | full aquifer<br>.03-.09<br>single level<br>.007 |

Note: n.r. means not reported and n.a. means not applicable

TABLE 2.3 Review of Field Tracer Tests in Crystalline Rock

| Reference | Type of Test | Medium | Tracer | Scale | Duration | Evaluations |
|---|---|---|---|---|---|---|
| Lewis et. al. (1966) | borehole dilution | crystalline | fluorescein dye | - | 0 days | seepage velocities; hydraulic conductivities |
| Webster et al. (1970) | two-well injection/ withdrawal | chlorite-hornblende schist & hornblende gneiss with some quartzite | tritium | 530 m | 2 years | dispersivity |
| Landstrom et al. (1970) | radial convergent (multiple well) | granite | $^{24}Na$, $^{75}Se$, $^{82}Br$, $^{85}Sr$, $^{99m}Tc$, $^{113}Sn$, $^{131}I$, $^{137}Cs$, $^{147}Nd$ | 51 m | 17 days | transit times; dispersion coefficients; retardation factor for $^{85}Sr$ |
| Calmels et al. (1979) Goblet et al. (1979) | radial convergent (multiple well) | granite | $Cl^-$, $I^-$, rhodamine WT, $NO_3^-$, $Sr^{2+}$, $Cs^+$, uranin | 12 m | | dispersiviites; fracture distribution coefficients; kinetic sorption & desorption coefficients |
| Schneider (1979) | radial divergent (multiple well) | granite | salt | 5 m | 35 minutes | transit time |
| Marine (1980) | in-hole | chlorite-hornblende schist & hornblende gneiss with some quartzite | $^{131}I$ | - | 9 days | positions of most transmissive fractures |
| Davison (1982) | two-well injection/ withdrawal | granite | $I^-$ | 10 m | 1 day | hydraulic diffusivity; Na/Ca sorption/desorption |
| Grisak et al. (1983) | radial convergent (multiple well) | granite | $SCN^-$, acetate, benzoate, KHP | 15 m | 1 day | transit times; relative tracer velocities |
| Novakowski et al. (1983a) | radial divergent (multiple well) | granite | $^{131}I$ | 11 m | 9 day | transit times; apertures; fracture interconnectivity |
| Novakowski et al. (1983b) | radial divergent (multiple well) | granite | $^{131}I$, $^{85}Sr$ | 11 m | 14 day | transit times; fracture retardation factor & fracture distribution coefficient for $^{85}Sr$ |

$c/\tau^2$). The total volume of host-rock pores with nominal diameters greater than approximately 0.1 μm, as well as the size distribution of these pores, can be measured with mercury-infusion porosimetry. Effective diffusion coefficients can be obtained using a diaphragm diffusion cell. Appropriate electrical measurements yield information on the tortuosity and constrictivity of the interconnected pore space in a rock. The validity of porosity, tortuosity, and constrictivity measurements can be checked directly in a qualitative way by optical and scanning electron microscopy (SEM). However, caution must be exercised in applying laboratory data on porosities and effective diffusion coefficients to natural rock formations, because these parameters tend to decrease systematically with increasing depth due mainly to the pressure exerted by the weight of overlying rock.

Practical applications of laboratory results on matrix diffusion are not necessarily straightforward. This is particularly true in the case of laboratory determinations of porosity, mainly because these data are influenced significantly by the pressure-temperature conditions (generally 0-2000 psi and 25°C) under which the measurements are made. In the natural or man-made subterranean environment to which the results of laboratory tests are ultimately intended to apply, the ambient pressure-temperature conditions will generally be different, and this can have significant effects on porosity. Specifically, pressure tends to increase with increasing depth due to the increasing load of overlying water (hydrostatic pressure), rock (lithostatic pressure), or water plus rock (a pressure intermediate between hydrostatic and lithostatic pressure), and these increases in pressure will tend to decrease the sizes of pores in a rock. On the other hand, due to geothermal gradients, temperature also increases with increasing depth, and increasing temperature

will tend to cause pores to expand, which counteracts the effect of pressure. Furthermore, because effective (matrix) diffusion coefficients for solutes are partly a function of the sizes of the pore in the rocks through which diffusion is occurring - specifically, matrix diffusion coefficients will tend to decrease with decreases in the sizes of the pore in a rock - it is to be expected that, in general, the rates at which solutes diffuse through rocks will diminish with increasing depth below the earth's surface.

There have been very few field tests of matrix diffusion in natural rock formations, and the only results that have been described in detail were obtained in an experiment performed in granite in the Stripa Mine in Sweden (Birgersson and Neretnieks, 1982). The following description of the field test is extracted from Blencoe and Grisak (1984).

The Stripa test was conducted at the working (heater) level in the Stripa Mine, approximately 360 m below the surface. A special drift had previously been excavated at this level for the purpose of performing various experiments, including a test of matrix diffusion in the granite near the end of the drift. In order to ensure that drift construction would have minimal effects on the results obtained in testing, smooth-wall blasting techniques were used to minimize fracturing of the granite. For the matrix diffusion experiment, access to essentially undisturbed rock was then accomplished by drilling a 14.5 cm diameter vertical hole 11 m into the rock floor at the end of the drift. [This distance is in accordance with the general rule of thumb that the depth (or length) of a borehole must be approximately 2 to 3 times the diameter of an access drift in order to reach rock that has not been significantly affected by the mining operations that were used to excavate the drift.]

After borehole drilling was completed, a second hole - 2 cm diameter and 2.0 m deep - was drilled in the rock at the bottom of the borehole. Next, a packer was emplaced at the bottom of the borehole to isolate the small volume of open space in the underlying 2 cm diameter hole. This open space was subsequently filled with an aqueous solution containing iodine, chromium-EDTA, and fluorescein (Uranine), which were the "tracers" for the matrix diffusion experiment. After injection, the pressure on the fluid was approximately 1 atm. (A fluid pressure in excess of 1 atm was not desired because "excess" pressure would have forced significant amounts of fluid into the granite, and this would have lead to erroneously high calculated matrix diffusion coefficients for the tracers.)

The tracer-bearing solution was allowed to remain in contact with the granite for three months, after which time the hole was overcored using a 13.2 cm I.D. core drill. The 2.5 m long cylinder of rock thus obtained was then sectioned to determine the distances that the tracers had diffused into the rock. [The effects of advective groundwater flow on tracer movement were negligible because the permeability (hydraulic conductivity) of the rock was measured and found to be between $10^{-12}$ and $10^{-13}$ m/s.]

Approximately 5-10% of the injection concentration of each tracer was found at a distance of 10.9 cm from the injection hole. The resulting calculated matrix diffusion coefficients for the tracers were determined to be approximately $1 \times 10^{-10}$ $m^2/s$, which is nearly an order of magnitude lower than corresponding values obtained from laboratory measurements. The systematic differences between the coefficients obtained from the field test and laboratory experiments have been ascribed to a lower

matrix porosity for the granite in the mine, this decreased porosity being attributable to the pressure exerted by the weight of overlying rock.

2.2.2.4 Retardation Term

Although the advection and dispersion terms are strongly scale dependent in their application, the reactive terms in the solute transport equation are somewhat more general in nature. The initial solute transport equation presented in this chapter can be simplified to some extent by assuming that the solute concentrations in the sorbed (S) and liquid (C) phases follow the linear relationship

$$S = K_d C \tag{2.26}$$

where $K_d$ is the proportionality constant referred to as the distribution coefficient. For cases where the partitioning of the solute can be adequately described by the distribution coefficient (i.e. linear adsorption isotherm and reversible adsorption without kinetic effects) the apparent velocity of the reactive solute $V_c$ is

$$V_c = \frac{\bar{V}}{1 + \frac{\rho_b K_d}{\theta}} \tag{2.27}$$

where the lower terms on the right hand side of the equation are referred to as the retardation factor R', expressed

$$R' = 1 + \frac{\rho_b K_d}{\theta} \tag{2.28}$$

This form of the volume-related retardation factor is based on the development for porous media, where reactions are expressed on the basis of the bulk density of the porous medium. It has been suggested (Burkholder and De Figh-Price,

1976) that, for fractured media, the retardation factor be expressed on the basis of surface areas rather than bulk densities, since the reactive solutes will be interacting with the fracture surfaces and, only through diffusion, with the bulk medium. Since there are reactions with the bulk medium (rock matrix) via the diffusive term, these should be taken into account when the retardation factor is employed. This can be done by coupling the fracture transport and matrix diffusion processes as presented in the early sections in this chapter. However, under certain conditions (i.e. high water velocity, small matrix diffusion component and relatively reactive fracture surfaces) the retardation factor $R_f'$ for transport in the fracture can be expressed

$$R_f' = 1 + \psi K_a \qquad (2.29)$$

where $K_a$ = $\dfrac{\text{mass of sorbed solute per unit surface area}}{\text{concentration of solute in liquid phase}}$

$\psi$ = surface area to fracture volume ratio, $A_f/v$

In order to relate the fracture retardation factor to the fracture flow parameters presented earlier, the limiting case of a planar fracture with smooth parallel sides is again involved. In this instance

$$R_f' = 1 + \frac{2K_a}{(2b)} \qquad (2.30)$$

However, it should be recognized, given the nonplanarity of actual fractures and the uncertainty in the amount of surface areas available for sorption, that a field tracer test would not likely provide data allowing separation of the parameters or terms within the retardation factor expression as presented in equations 2.29 and 2.30.

Mukhopadhyay and Hadermann (1982) derived the fracture retardation factor from a continuity equation for solute transport in fractured media. They showed theoretically, that with a sufficiently large REV, the $R_f'$ term correctly accounts for all fracture and matrix pore volumes. Although this is theoretically correct, the concept does not readily lend itself to field evaluation. Mukhopadhyay and Hadermann (1982) also showed the manner in which the retardation term developed for a fractured medium can be transferred to an effective porous medium conceptualization when there is a system of dense fractures.

The reactive term, or the geochemical retardation term, is employed in many instances as a simple retardation factor, with little or no thermodynamic significance. For instance Pickens et al. (1981) measured retardation factors for $^{85}Sr$ in the sand aquifer at CRNL by employing the radial tracer injection system illustrated in Figure 2.37. In this instance the comparison between field and laboratory measured values was very good. A retardation factor of about 55 for $^{85}Sr$ in crystalline rock (Figure 2.39) has also been measured at the CRNL rock site (Novakowski et al., 1983).

Although many thermodynamically-based theoretical developments can be derived to develop the geochemical retardation term, one of the most straightforward methods of determining the transport rates for reactive solutes is to directly measure the solute of interest in field tests under appropriate field conditions. A review of reactive field tracer testing techniques is provided in GTC (1983), while a summary of reactive tracer tests which have been performed is presented in Table 2.4 (adapted from Pickens et al., 1981).

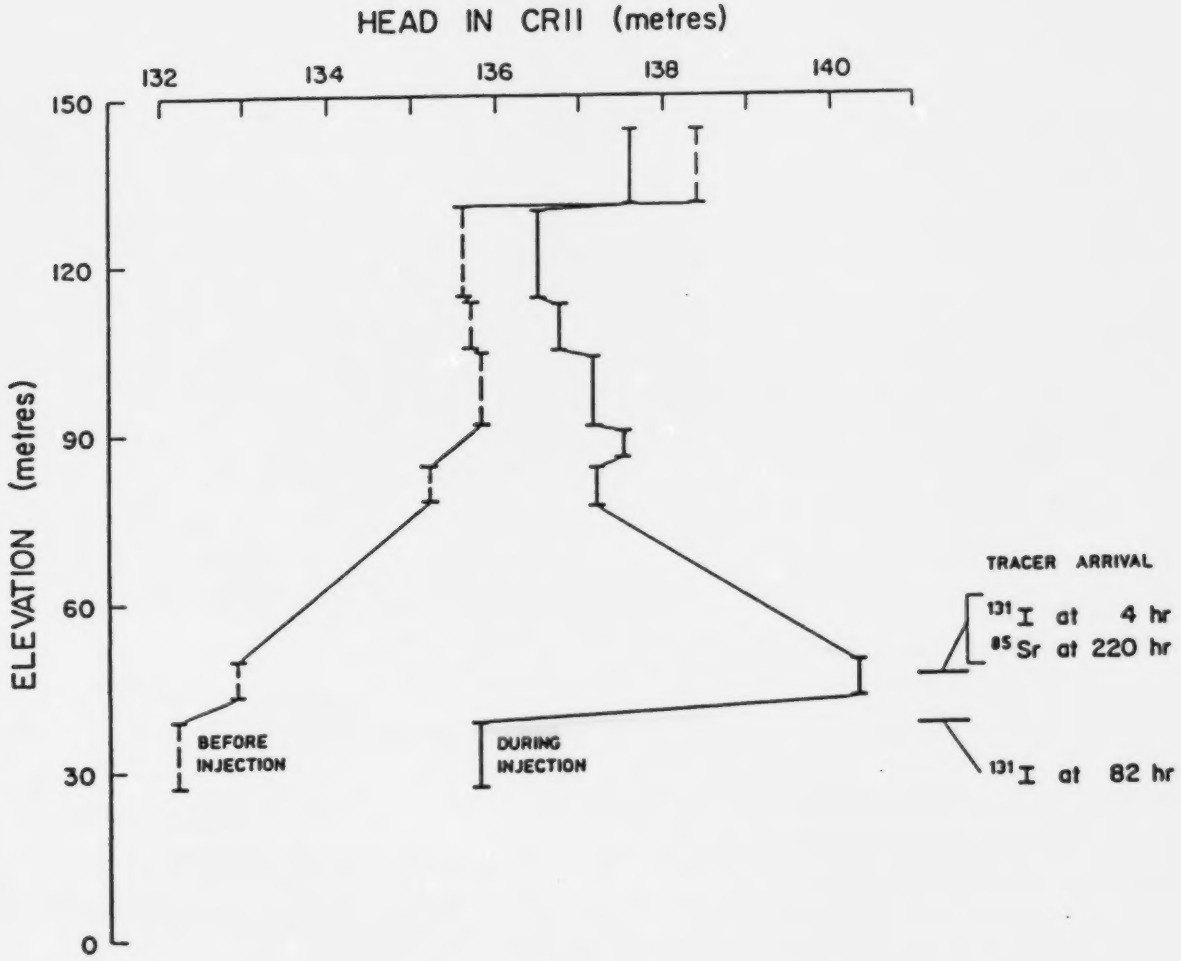



Figure 2.39 Pressure Profiles and Tracer Arrival at CRNL Rock Tracer Site

TABLE 2.4 Review of Field Reactive Tracer Tests

| Reference | Type of Test | Medium | Tracers | Evaluations |
|---|---|---|---|---|
| Ewing (1959) | radial injection | unconsolidated gravel and sand | $^{89}Sr$ | |
| Kaufman (1960) | two-well injection and withdrawal plus intermediate sampling wells | consolidated gravel and sand | tritium, $^{89}Sr$ | equilibrium constant; rate coefficient |
| Inoue and Kaufman (1963) | central injection well with four outer withdrawal wells | clay, sand and gravel | $^{89}Sr$ | distribution coefficient |
| Parsons (1963) | natural gradient | sand | tritium, fluorescein, $^{89}Sr$, $^{60}Co$, chloride, organic dye | relative velocities compared well breakthrough curves for withdrawal phase |
| Landstrom et al. (1978) | multilevel pulse (central withdrawal well with tracer pules injection to outer wells) | fractured granitic rock | selenium, technetium tin, cesium, iodine, neodymium, strontium, bromide, sodium | dispersion coefficients; retardation factor for $^{89}Sr$ |
| Roberts et al. (1978) | radial injection | gravel; coarse to fine | wastewater treatment plant effluent: chloride, ammonia, iron, dissolved oxygen, total organic carbon, major ions, trace metals | compared breakthrough response of various constituents |
| Calmels et al. (1979), Goblet et al. (1979) | multiwell pulse (central | fractured granitic rock | $Cl^-$, $I^-$, Rhodamine-WT, $NO_3^-$, $Sr^{2+}$, $Cs^+$ | dispersivities; fracture distribution coefficients; sorption and desorption coefficients |
| Pickens et al. (1981) | single-well injection withdrawal with multisamplers at radial distances | sand | $^{131}I$, $^{85}Sr$ | distribution coefficient, dispersivity |
| Novakowski et al. (1983) | central injection packed off zone with multilevel monitoring in radially spaced boreholes | crystalline rock | $^{131}I$, $^{85}Sr$ | velocities, retardation term |

2.2.2.5 Effective Porosity
Matrix Porosity

The effective porosity of samples of porous geologic media can be determined under field or laboratory conditions. For situations involving fractured media, tracer tests are an effective means for determining porosity. For example, in two-well tracer tests, porosity can be used as the calibration parameter in fitting calculated-to-measured tracer concentrations at the withdrawal well. The matrix porosity of rocks is generally obtained under laboratory conditions using nitrogen adsorption methods, mercury-infusion porosimetry (Gregg and Sing, 1967) successive granulation (Norton and Knapp, 1977) and direct examination by optical microscopy and/or SEM. Among these methods, mercury infusion porosimetry is particularly useful because it can be used to measure not only the effective porosity of a rock sample, but also the size distribution of pores and the total volume of the sample. When combined with the results of grain density determinations, measurements of total sample volume obtained from mercury-infusion porosimetry can be used to calculate total porosities.

2.2.2.6 Radionuclide Source Terms

The radionuclide source terms are dependent upon such factors as concentration, the radionuclide decay chains, solubility rates and leach rates (waste emplacement containers and backfill). Source terms are usually estimated or derived from data on solubility and leach rates generally determined under laboratory conditions with variables being pressure, temperature, water chemistry and container material.

2.2.2.7 Radioactive Decay Constant

The radioactive decay constants are known for the radio-

isotopes of interest. Chain decay is usually incorporated into modeling studies with analytical or numerical techniques (e.g., Lester et al., 1975; INTRACOIN, 1984).

## 2.3 Heat Transport

Heat transport in hydrogeologic systems is governed by the processes of conduction through the solid/liquid medium and convection by flowing ground water. Ground water usually does not flow at sufficiently high rates to markedly affect heat transport, however heat does have a significant effect on ground water flow patterns, due to the thermal properties of water. Heat sources, such as radioactive waste repositories in ground water systems, can cause significant flow field effects which may be quite extensive compared to the size of the system.

### 2.3.1 Governing Equations

The heat transport equation describing the transient temperature distribution of a fluid in a saturated porous medium can be written in two dimensions (as adapted from Bear, 1972)

$$\begin{aligned} L(T) = & \left(\rho\theta\ C_{vw} + (1-\theta)\ \rho_s\ C_{vs}\right)\frac{\partial T}{\partial t} \\ & - \frac{\partial}{\partial x}\left[\left(\theta K_{xx}^{\ w} + (1-\theta)\ K_{xx}^{\ s}\right)\frac{\partial T}{\partial x}\right. \\ & \left. + \left(\theta K_{xz}^{\ w} + (1-\theta)\ K_{xz}^{\ s}\right)\frac{\partial T}{\partial z}\right] \\ & - \frac{\partial}{\partial z}\left[\left(\theta K_{zx}^{\ w} + (1-\theta)\ K_{zx}^{\ s}\right)\frac{\partial T}{\partial x}\right. \\ & \left. + \left(\theta K_{zz}^{\ w} + (1-\theta)\ K_{zz}^{\ s}\right)\frac{\partial T}{\partial z}\right] \\ & + \rho\ C_{vw}\left(q_x\frac{\partial T}{\partial x} + q_z\frac{\partial T}{\partial z}\right) \\ & + Q^w + Q^s \\ = & \ 0 \end{aligned} \qquad (2.31)$$

where T is the temperature, $C_{vw}$ is the heat capacity of the liquid phase at constant volume per unit mass, $\rho_s$ is the density of the solid phase, $C_{vs}$ is the heat capacity of the solid phase at constant volume per unit mass, $K_{xx}^w$, $K_{xz}^w$, $K_{zx}^w$, and $K_{zz}^w$ are the components of the thermal dispersion tensor (which includes the effects of mechanical dispersion and thermal conduction) of the liquid phase, $K_{xx}^s$, $K_{xz}^s$, $K_{zx}^w$ and $K_{zz}^s$ are the components of the thermal conductivity tensor of the solid phase, and $Q^w$ and $Q^s$ are heat liquid and solid point source or sink functions. This form of the equation includes dispersion (both conduction in liquid and solid phases and mechanical dispersion in the liquid phase), advection (both free and forced convection) and a heat source or sink. The heat liquid source or sink function $Q^w$ is related to the liquid source or sink function $Q^p$.

Heat transport by advection is usually considered slow enough to allow the temperature of the solid and liquid phases to be assumed to be equal at any location in the flow domain. It is conceivable that this assumption could introduce errors in a fractured-media flow system which exhibits high water velocities and few fractures.

The thermal dispersion tensor for the liquid phase in Equation (2.31) includes the effects of mechanical dispersion and thermal conduction. It is assumed that it can be written in a similar fashion to that suggested by Reddell and Sunada (1970) for solute transport in an isotropic medium. The components are written:

$$\theta K_{xx}^{w} = \rho C_{vw} \left( \frac{\alpha_L q_x^2}{q} + \frac{\alpha_T q_z^2}{q} \right) + \theta K^{w*} \qquad (2.32a)$$

$$\theta K_{zz}^{w} = \rho C_{vw} \left( \frac{\alpha_T q_x^2}{q} + \frac{\alpha_L q_z^2}{q} \right) + \theta K^{w*} \qquad (2.32b)$$

$$\theta K_{xz}^{w} = \theta K_{zx}^{w} = \rho C_{vw} \left( \alpha_L - \alpha_T \right) \frac{q_x q_z}{q} \qquad (2.32c)$$

where $\alpha_L$ and $\alpha_T$ are the longitudinal and transverse dispersivities, respectively, of the medium, q is the resultant Darcy velocity of the liquid phase and $K^{w*}$ is the thermal conductivity of the liquid phase. The components of the thermal conductivity tensor of the solid phase are assumed to be coincident with the Cartesian directions and hence:

$$K_{xz}^{s} = K_{zx}^{s} = 0 \qquad (2.32d)$$

A number of researchers (Mercer et al, 1975; Bear 1972; Green et al. 1964; Weber 1975; Rubin, 1974, Andrews, 1978; Popadopulos et al., 1978) assumed that heat dispersion includes both a liquid velocity-dependent component and a thermal conduction component. In contrast, several other papers (Sorey, 1975; Faust and Mercer, 1977a; Lippman et al, 1977; Faust and Mercer 1977b) considered the heat dispersion coefficient as a scalar for the liquid-solid matrix.

The initial conditions necessary for the solution of Equation (2.31) are:

$$T(x,z,0) = \tilde{T}_o(x,z) \qquad (2.33)$$

where $T_o$ are the initial prescribed temperatures. The Dirichlet boundary conditions can be expressed as:

$$T(x,z,t) = \tilde{T}(x,z,t) \text{ on } \Gamma_6 \qquad (2.34a)$$

and the Neumann boundary conditions can be expressed as:

$$\left[ \left( \theta K_{xx}^{w} + \theta K_{zx}^{w} + (1-\theta)\, K_{xx}^{s} \right) \frac{\partial T}{\partial x} + \left( \theta K_{xz}^{w} + \theta K_{zz}^{w} + (1-\theta)\, K_{zz}^{s} \right) \frac{\partial T}{\partial z} \right] \bar{n} = -q_n^{T} \quad \text{on} \quad \Gamma_7 \tag{2.34b}$$

where $\tilde{T}$ are prescribed temperatures, $\Gamma_6 + \Gamma_7 = \Gamma$, the total boundary system, and $q_n^T$ is the prescribed heat flux across a Neumann boundary.

As noted earlier, to apply the heat transport equation, it is usually assumed that heat transport by advection is slow enough to allow the temperature of the solid and liquid phases to be assumed to be equal at any location in the flow domain. This assumption can be considered correct on first approximation at the regional scale, but is erroneous at the single fracture scale. It can be shown, however, that heat conduction is the dominant heat transport mechanism at most conceivable ground water velocities, and although water velocity may be high in a fracture, the adjacent rock matrix acts much like a heat sink, thereby limiting the heat transport mechanisms away from a source to heat conduction.

### 2.3.2 Parameter Evaluation (Data Required to Quantitatively Evaluate Heat Transport)

Most of the heat transport parameters and required data are obtained from laboratory measurements on core or other samples or from functional relationships between water properties which are listed below.

2.3.2.1 Heat Capacity

The heat capacity parameter is required for solid and liquid phases. Laboratory measurements are generally utilized.

2.3.2.2 Thermal Conductivity

Thermal conductivity values are required for solid and liquid phases. Laboratory measurements are generally utilized.

2.3.2.3 Water and Solid Phase Density

Functional relationships between density, temperature and total solute concentrations, are required.

2.3.2.4 Thermal Dispersion Coefficient

Mechanical dispersion component for heat transport is usually considered equivalent to the mechanical dispersion component for solute transport. The thermal dispersion coefficient includes both conduction in the liquid and solid phases and mechanical dispersion in the liquid phase. Numerical modeling studies of field thermal transport use either the conduction and mechanical dispersion approach noted above or use a lumped thermal disperson coefficient that is a scaler for the liquid-solid matrix. There is a lack of uniformity in this regard; however, both approaches have yielded results representative of the field situations being investigated.

# 3.0 RESEARCH PROGRAM AND DATA ASSIMILATION

A hypothetical research and investigative program is proposed in this section for the purposes of illustrating the hydrogeologic methods and research techniques that the authors of this report consider applicable and suitable for site characterization studies within a regional system (Figure 3.1).

The proposed program is generic in nature in that it does not attempt to establish a final assessment of waste emplacement consequences (i.e., dose calculations) but rather attempts to establish the hydrogeologic regime in sufficient detail to allow quantitative input of the hydrogeologic data to such an assessment. A listing of data acquisition methods currently employed in several national radioactive waste programs is also provided in Appendix 4.

## 3.1 Regional Characterization

The starting point in a site characterization is to establish the nature of the larger scale regional hydrogeologic system within which the site is located. The geologic and structural framework is the starting point. The regional hydrogeologic system is likely bounded by major topographic highs and primary drainage. For instance, the Western Canada Sedimentary Basin is considered as a regional hydrogeologic flow system, with major recharge in the Foothills and Front Range of the Rockies and major discharge along a line between Great Slave Lake and Lake Winnipeg (Figure 3.2). Secondary recharge to the regional system occurs in the area of isolated topographic highs such as the Cypress Hills, Cameron Hills and Clear Hills.

In the Canadian Shield the size of the regional flow systems is likely to be somewhat smaller because the relief is lower. For instance the flow system in the Atikokan Research Area study is

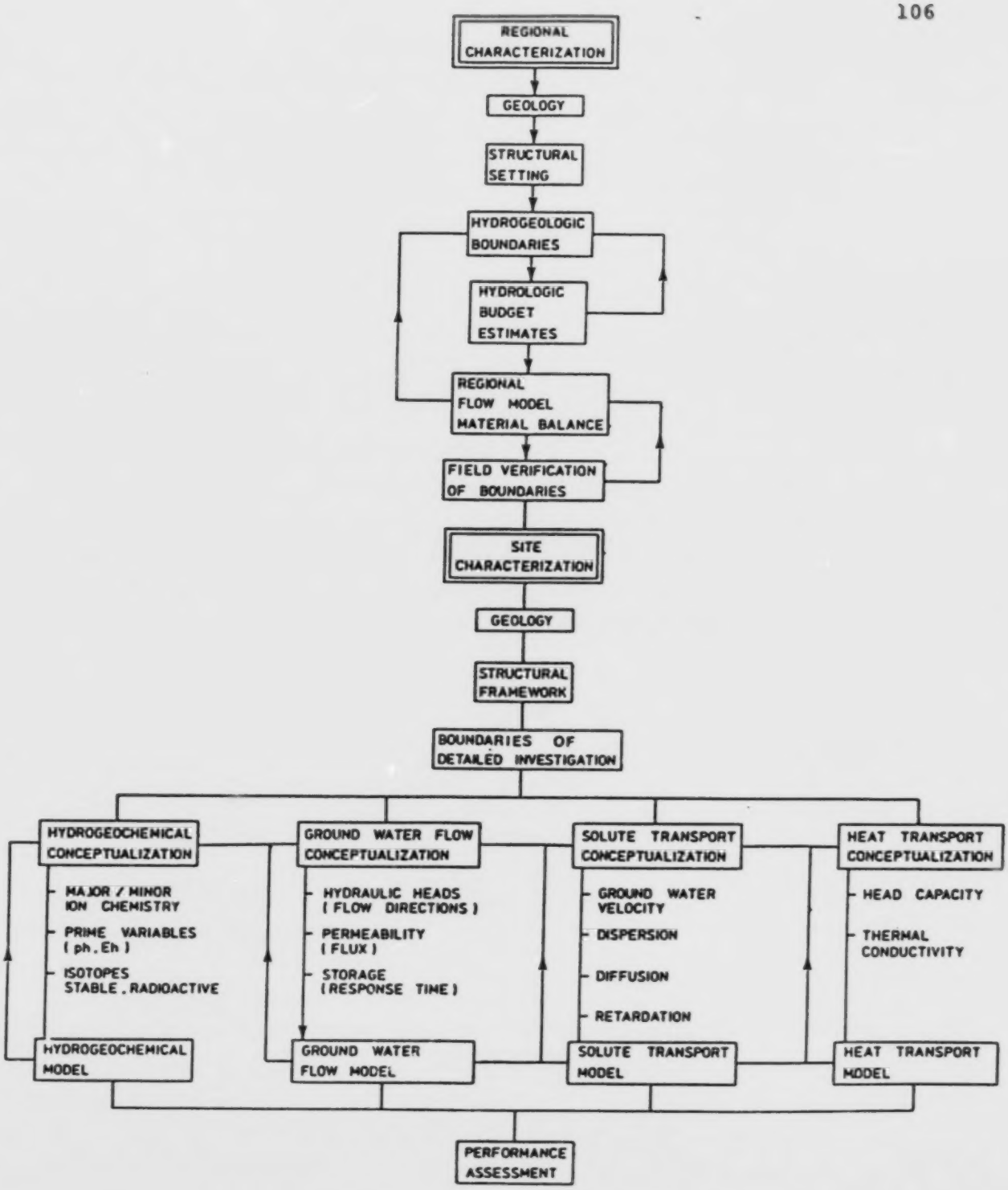



Figure 3.1 Hydrogeologic Methods for Regional and Site Characterization

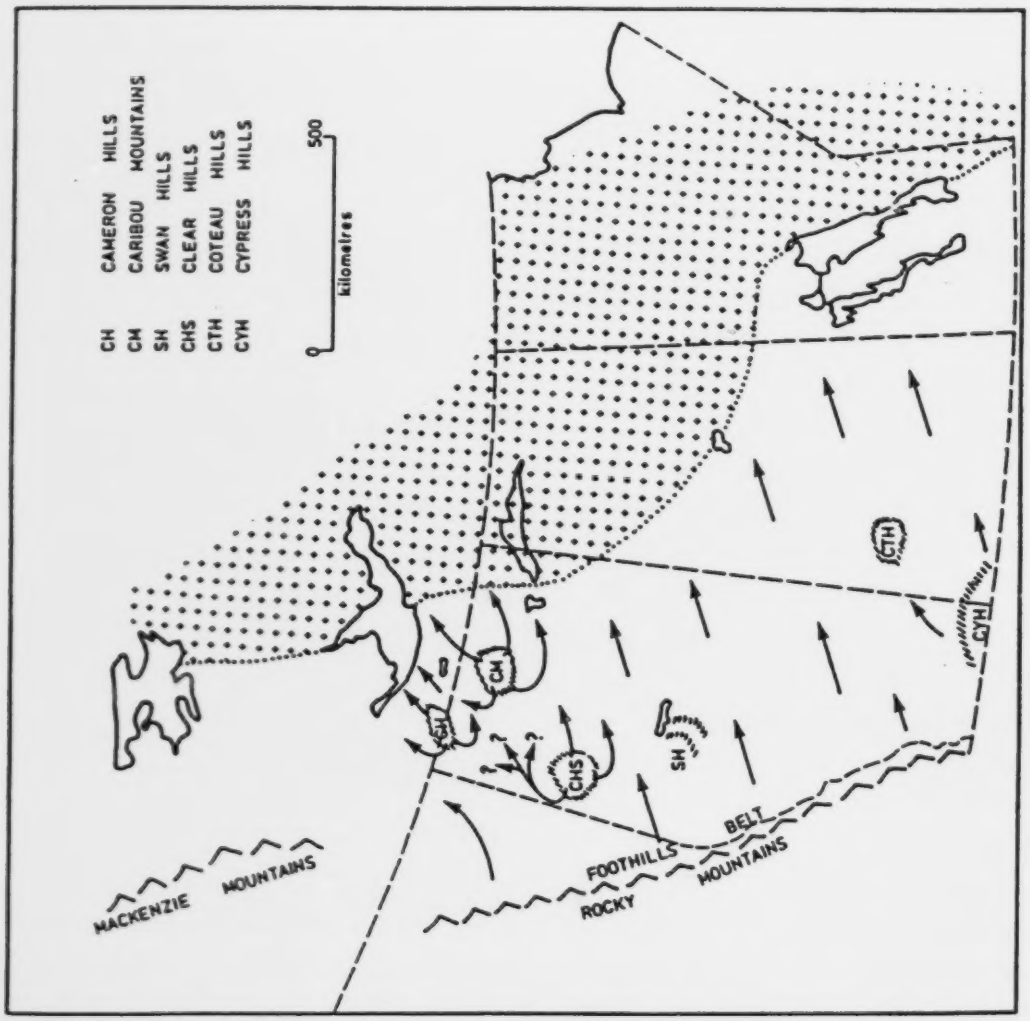



Figure 3.2 Schematic of Western Canada Sedimentary Basin ground water flow directions (from GTC, 1983)

hypothesized to be bounded by the topographic highs and lows within a few tens of miles of the pluton.

Regional flow system interpretation is usually based more on the principles of ground water movement rather than direct measurement of the ground water system. In recent years it has become evident that three-dimensional regional ground water flow models can serve as very useful interpretive tools. The flow models, although only as accurate as the input data, can provide a rough approximation by conducting a material balance within a hypothesized system. If the regional ground water flow system is hypothesized as bounded by major topographic highs and drained by major rivers, a rough material balance can be conducted, whereby a percentage of precipitation as recharge is balanced against the 'baseflow' (non-runoff-related) discharge in the river. It is possible to determine if the approximate size of the system under consideration can be related to the apparent ground water removal from a region. These types of studies are referred to as 'hydrologic or water budget' studies. They represent one of the earliest steps in a hydrogeologic investigation.

The next step is to verify the water budget studies with actual measurements of baseflow discharge to streams and rivers, and perhaps direct infiltration measurements, using lysimeters or controlled recharge studies.

Verification of the boundaries of the system are considered to be one of the most critical aspects of a hydrogeologic investigation. If the boundaries of a system are adequately defined, then the actual ground water flow regime is determined by the permeability configuration internal to the bounded system. One of the most straightforward boundaries that can be identified is the water table, since it requires only direct measurements (with wells/piezometers) in shallow boreholes and limited geologic interpretation (compared to deeper flow regimes).

The peripheral boundaries of the system are more difficult to verify than the water table, however it is not an exceptionally difficult task to establish the hydraulic head profile along the hypothesized peripheral boundaries. The lower boundary of deep hydrogeologic systems is more problematic. It is often assumed that permeability decreases with depth and that flow at great depth is sufficiently low volumetrically that there is a lower limit to the 'active' flow regime.

It is the author's opinion that a decrease in permeability with depth is an extreme oversimplification, although the fact of greater overlying stress causing decreased pore space is obvious. Based on the results of deep testing in several national radioactive waste programs, it can be stated that permeability decreases with depth, except where it doesn't. High permeability fracture zones or discreet fractures seem to be, if not common, at least not unexpected at depth.

Once the hydrogeologic boundaries of the regional system have been hypothesized and rough hydrologic budget estimates indicate some measure of consistency in the area selected and the potential recharge/discharge, the flow model should be used to provide an approximate material balance for the system. The final step in the regional characterization is to verify the nature and existence of the system boundaries. This can be done both by direct measurement and by inference from other data sets (for instance the hydrogeochemistry).

On satisfactory completion of regional characterization and verification, site characterization studies can then be incorporated into the regional conceptualization. It is not necessary to complete the regional characterization before commencing the site characterization, since both data sets should be complementary. It may also be necessary to iterate and alter previous conceptualizations, either regional or site, on

acquisition, interpretation and assimilation of new data. This should be considered simply as part of the investigation process and as a continued refining of the interpretation.

## 3 2 Site Characterization

Site characterization studies (Figure 3.1) are essentially those aimed at providing sufficient quantitative data on which to base performance assessments (performance assessment is defined as the quantitative prediction of the long term performance of a radioactive waste isolation system after the waste has been placed within a geologic repository and the repository has been closed).

The following paragraphs will simply identify the necessary data and indicate the preferred investigation methodology which could be employed to acquire the data. Other site assessment techniques, including geophysical techniques, are considered to be useful for confirmatory or correlative efforts, but not particularly applicable for data aquisition leading to a quantitative performance assessment.

**Ground Water Flow**

- scale definition required to apply REV concepts

- Hydraulic Heads

  - measured with piezometer nests in overburden
  - single packer or through the bit shut-in measurements immediately after an interval is intersected in rock
  - long term multilevel pressure monitoring (baseline) in boreholes on completion of drilling and open hole testing

- three-dimensional hydraulic head patterns derived from data

- Permeability

  - single well and multiple well pumping/slug tests in overburden
  - permeability profile over complete open borehole in rock
  - multi-hole tests with multiple completions
  - develop velocity field from permeability and head data

- Storage

  - from multi-hole tests in rock

- Interconnectivity

  - fracture interconnectivity with respect to pressure should be determined using multi-well tests

- Ground Water Flow Model

  - three-dimensional flow model developed of the area using local boundaries or those derived from situation within regional system
  - iterate with field program: should occur in parallel with field program

**Solute Transport**

- scale definition required to apply REV concepts

- Ground Water Velocity

  - the velocity field developed from the flow regime should be verified with tracer tests;

- point measurements (borehole dilution) and multi-borehole tests

- Dispersion

  - dispersion characteristics and sampling methodology consistent with scale definition should be determined using tracer methods

- Interconnectivity

  - fracture interconnectivity suggested by flow tests should be verified for solute using multi-well tests

- Diffusion

  - in fractured media diffusion characteristics of matrix should be determined using direct laboratory methods

- Retardation

  - retardation factors for elements of interest should be determined directly by field reactive tracer tests, using either the element itself, a short-lived isotope or a stable analogue.

- Solute Transport

  - a solute transport model consistent with the velocity field should be developed. The model development should be an interactive process with the parameter measurement and the ground water flow conceptualization activities discussed above.

**Heat Transport**

- Heat Capacity

  - the heat capacity for both the rock and the fluid (ground water) are straightforward laboratory measurements (if the existing data in the literature are deemed inadequate).

- Thermal Conductivity

  - the thermal conductivity of the fluid and media are laboratory determined parameters

- Thermal Expansivity

  - the thermal expansivity of water and its pressure, temperature and dissolved solids functional relationships are laboratory measured properties

- Heat Transport Model

  - a heat transport model, consistent with the flow and solute transport models should be developed in the same iterative data/model/data manner as the flow and solute models
  - site specific field tests are not considered necessary

**Hydrogeochemistry**

- Major/Minor Ion Chemistry

  - should be determined in laboratory on samples collected and preserved according to accepted protocol
  - samples should be "representative" of system ground water

- Prime Variables

  - prime variables such as pH/Eh should be measured in the field (techniques not discussed in this report)

- Dissolved Gases

  - measured in laboratory on representative samples collected in pressure vessels and retained at in-situ pressure

- Isotopes - Stable/Radioactive

  - stable and radioactive isotope samples should be obtained using established protocol

- Hydrogeochemical Model

  - hydrogeochemical models, both equilibrium and reaction path types, should be developed and be shown to be consistent with the other models and data concerning the hydrogeologic system

**Performance Assessment**

A quantitative assessment of the long term performance of the site as a radioactive waste repository is then conducted by assembling the sub-models described above into a performance assessment framework. The performance assessment results should be those on which the suitability of a site is determined by the regulatory agencies.

## 4.0 RECOMMENDED RESEARCH

The following areas of hydrogeologic research are perceived by the authors as areas that could provide considerable input to the development of regulatory attitudes and criteria:

- Regional Characterization Studies
  - quantitative
  - emphasis on boundary definition
- Assessment of the transferability and significance of REV interrelationships between flow, solute transport, geochemistry and heat transport
- Improvements in field measurement and interpretive techniques for pressure, permeability and hydrogeochemistry in low permeability media
- Development of a comprehensive stochastic statement of the problem of flow and transport in geologic media
- Development of a comprehensive risk assessment model for hydrogeologic systems associated with repositories
- Development of a fully coupled ground water flow/hydrogeochemical solute transport mathematical model

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# APPENDIX 1

## Permeability of a Fractured Zone

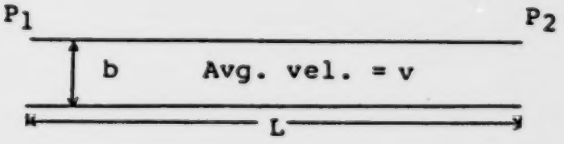
## Permeability of a Fractured Zone

This discussion is written to assist in understanding the results of the flow tests in crystalline rock in terms of the fracture dimensions.

An isolated interval may or may not contain any fractures. In this Appendix, permeability, conductivity and transmissivity of a fractured zone are calculated given the fracture distribution number and sizes. First, hydraulic conductivity and transmissivity of a single fracture are calculated, and then the results are correlated to multiple fractures over a certain thickness.

The basic assumptions are that (1) laminar flow exists in all the fractures, and (2) fractures are uniform and can be modeled individually as a set of parallel plates.

### Single Fractures




By comparing average fluid velocity in terms of pressure drop over a given length, and Darcy velocity (Annex 1), single fracture

$$\text{permeability } k_f = \frac{b^2}{12} \qquad (1)$$

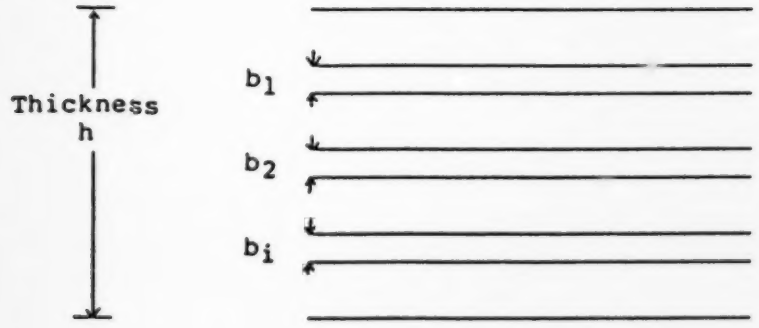
where b is the fracture width, and

$$\text{hydraulic conductivity } K_f = \frac{(\rho g)}{(\mu)} \frac{b^2}{12} \qquad (2)$$

where $\rho$ = fluid density, and
$\mu$ = fluid viscosity.

Therefore
transmissivity $T_f = \frac{\rho g}{12\mu} b^3$ (3)

Multiple Fractures




Assuming that matrix permeability is negligible relative to the fracture permeabilities.

Transmissivity of the formation of thickness h is

$$T = \sum_{i=1}^{N} K_i b_i = T_i \quad (4)$$

where N = Number of fractures over the formulation thickness h or

$$T = \frac{\rho g}{12\mu} \sum_i b_i^3 \quad (5)$$

if the fractures are equal size, i.e. $b_i = b$, then

$$T = \frac{\rho g N b^3}{12\mu} \quad (6)$$

This is the quantity measured in the field.

Average conductivity of the formulation

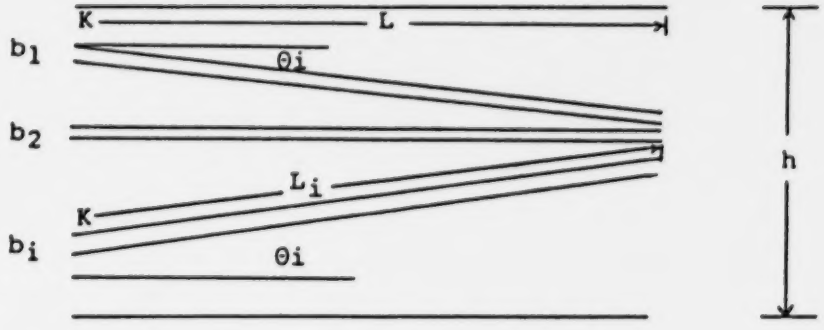
$K = T/h$

$$= \frac{\rho g}{12 \mu h} \sum_i b_i^3 \qquad (7)$$

and average permeability

$$k = \frac{1}{12 h} \sum_i b_i^3 \qquad (8)$$

Fractures Not Parallel to the Formation Axis




Then we have

$$T = \frac{\rho g}{12 \mu} \sum_i b_i^3 \cos \theta_i \qquad (9)$$

$$K = \frac{\rho g}{12 \mu h} \sum_i b_i^3 \cos \theta_i \qquad (10)$$

$$k = \frac{1}{12 h} \sum_i b_i^3 \cos \theta_i \qquad (11)$$

ANNEX 1

Pressure drop for flow through a set of parallel plates over length L is given as:

$$\Delta p = p_1 - p_2 = \frac{\rho f L v^2}{2b} \qquad (1)$$

where $\rho$ = Fluid density
and $f$ = Friction factor

In terms of fluid head,

$$\Delta p = \rho g \Delta \Phi \qquad (2)$$

$$\therefore \Delta \Phi = \frac{f L v^2}{2 g b} \qquad (3)$$

For laminar flow through parallel plates,

$$f = \frac{24}{R_e} \qquad (4)$$

where $R_e$ = Reynolds number

$$R_e = \frac{b v \rho}{\mu} \qquad (5)$$

where $\mu$ = Fluid viscosity

Substituting Equations (4) and (5) in (1),

$$\Delta p = \frac{\rho L v^2}{2b} \cdot \frac{24 \mu}{b v \rho}$$

$$= \frac{12 L v \mu}{b^2}$$

or

$$v = \frac{(b^2)}{(12\mu)} \frac{(\Delta p)}{(L)} \tag{6}$$

Alternatively

$$v = \frac{(\rho g}{(\mu} \cdot \frac{b^2)}{12)} \frac{(\Delta \Phi)}{(L)} \tag{7}$$

Darcy's law can be written as

$$u = \frac{k_f}{\mu} \frac{(\partial p)}{(\partial x)} \tag{8}$$

where $u$ = Darcy velocity,
$k_f$ = Permeability of a single fracture

Assuming that the single fracture porosity is 100%, $v = u$.

In one-dimension, under steady-state conditions,

$$\frac{\partial p}{\partial x} = \frac{\Delta p}{L}$$

$$v = \frac{k}{u} \frac{\Delta p}{L} \tag{9}$$

In terms of fluid head,

$$v = \frac{\rho g}{\mu} k_f \frac{\Delta \Phi}{L}$$

$$= K_f \frac{\Delta \Phi}{L} \tag{10}$$

where $K$ = Hydraulic conductivity of a single fracture.

Comparing Equations (6) versus (9), and (7) versus (10), we have

$$k_f = \frac{b^2}{12} \tag{11}$$

and

$$K_f = \frac{\rho g\ b^2}{12\ \mu} \tag{12}$$

Transmissivity of a single fracture

$$T_f = K_f\ b$$

$$T_f = \frac{\rho g b^3}{12\ \mu} \tag{13}$$

Inclined Fractures

Where the fracture is not horizontal, the fracture length L; and the formation length $L_i$ are related as

$$L = L_i \cos \theta_i \tag{14}$$

Equation (6) changes to

$$v_i = \frac{b_i\ ^2}{12\ \mu} \cdot \frac{\Delta p}{L_i}$$

$$= \frac{b_i\ ^2 \cos \theta_i}{12\mu} \cdot \frac{\Delta p}{L} \tag{19}$$

or

$$k_i = \frac{b^2 \cos \theta_i}{12} \tag{20}$$

and

$$K_i = \frac{\rho g}{12 \mu} b^2 \cos \theta_i \qquad (21)$$

$$T = \frac{\rho g}{12 \mu} \sum_{i=i}^{N} b_i^3 \cos \theta_i \qquad (22)$$

$$k = \frac{1}{12 h} \sum_{i=i} b_i^3 \cos \theta_i \qquad (23)$$

If the fractures are not smooth, and Equation (4) is not valid, appropriate expression for the friction factor can be substituted, and expressions for permeability, conductivity and transmissivity can be derived in an identical manner.

APPENDIX 2

# Diffusion in Geologic Media

GTC

## Diffusion in Geologic Media

Diffusion in geologic media is a simple extension of diffusion principles, which apply to aqueous dissolved species, to the geologic framework. Based on studies of salt migration across animal tissue membranes, Fick (1855) formulated an empirical relationship between the flux (F) of a dissolved species, the area (A) across which the flux takes place, the concentration gradient (dC/dy) in a specific direction and a coefficient of proportionality (D, the diffusion coefficient)

$$F = AD \frac{dC}{dy} \qquad (1)$$

or in unit area and unit time form with a negative sign to indicate movement towards lower concentrations.

$$F = -D \frac{dC}{dy} \qquad (2)$$

Equation (2) is known as Fick's first law. The original postulation (Fick, 1855) was based on analogies with heat conductance and electrical conductance.

The driving force for diffusion is Brownian movement, or the random thermally induced movement of molecules, which was quantitatively described by Einstein (1905). Brownian movement, or diffusion in the direction of lower concentration, is caused by the difference in the number of impacts received by a molecule (dissolved species) from the upgradient and downgradient sides. Diffusion is clearly not along a single axis but occurs in three dimensional space in response to the concentration gradient. By considering the definitions of Brownian movement in an aqueous solution, it is obvious that the coefficient of proportionality,

D, is a function of temperature, pressure, concentration, molecular size and other dissolved components in the aqueous solution.

For practical purposes the pressure dependence is insignificant, with D changing minimally over the pressure range of 1 to 1000 atmospheres. Temperature dependence is more significant and can be explained along with the molecular size dependence by the Stokes-Einstein relationship (Lerman, 1979, p. 86)

$$D = \frac{KT}{6\pi ur} \qquad (3)$$

where k = Boltzman constant
T = temperature
u = viscosity of the fluid
r = radius of diffusing particle

The effects of concentration and multicomponent solutions have been addressed by a number of authors and summarized by Lerman (1979). The concentration effect on a single component solution is in the range of 0-40% (Lerman, 1979). In multicomponent solutions the effects on D are complicated by the variations of ionic strengths of individual species. However, the off diagonal or cross diffusion coefficients (i.e., $D_{ij}$ where $i \neq j$) appear to be most affected in simple two component solutions (Lerman, 1979).

For practical purposes diffusion coefficients are considered as constants in any particular environment, providing the pressure, temperature and total ionic strength of an aqueous solution remain relatively constant. In geologic repository environments this assumption may be invalid with respect to the thermal effect in the near-field, however in the general case of transport through any particular hydrogeologic environment the aqueous

diffusion coefficient for a single dissolved radionuclide is likely relatively constant.

While Fick's first law stipulates the flux relationships, then conservation of mass requires that any change in the flux between y and y+ Δy must be accounted for by a time rate of change in concentration within the distance y and y+ y. In diffferential notation

$$\frac{\partial C}{\partial t} = - \frac{\partial F}{\partial y} \qquad (4)$$

where t = time
C = concentration

and the partial derivatives are used because concentration C and flux F are functions of distance and time. If the system is at steady state, then

$$\frac{\partial C}{\partial t} = 0 \qquad (5)$$

and F is constant, which means that the concentration gradient must be maintained to retain a constant nonzero flux (these conditions will manifest as boundary conditions in later discussions). Substitution of equation (2) into equation (4) yields Fick's second law of diffusion

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial y}\left(D \frac{\partial C}{\partial y}\right) \qquad (6)$$

which relates the rate of change of concentration to the distance, the diffusion coefficient and the concentration gradient, or to the second derivative of concentration with respect to distance at a fixed point in the y direction.

Thus far the development of diffusion concepts has been concerned only with aqueous dissolved species. However, matrix diffusion occurs as a porous media process; that is, the cross-sectional area through which diffusion occurs is only that made available by the pore spaces or porosity ($\theta_m$) in the matrix. In addition, the diffusion is not necessarily linear in space, but may take place through rather tortuous flow paths which provide the pore space interconnectivity. This extended path length is referred to as the tortuosity ($\tau$), or the relationship between actual flow path length and linear distances between points. Bear (1972) discusses tortuosity and presents the development of the relationship between tortuosity and the formation factor, or electrical resistivity characteristic of porous media.

In addition to the effects of porosity and tortuosity on solute diffusion in porous media, there are additional effects most likely due to pore geometry. These effects are usually incorporated into models of porous media diffusion in the factor referred to as the constrictivity (c). The constrictivity accounts for variable aperture pores. This term has been considered by Melnyk (1983) to be more related to the transient diffusion process rather than the steady-state process. The constrictivity factor is usually included as a factor which modifies the aqueous or free water diffusion coefficient, $D_o$, when describing diffusion in a porous medium.

Walter (1982) discusses the relative effects of $\theta$, $\tau$ and c and presents the relationship between the diffusion coefficient in porous media ($D_p$) and the free water diffusion coefficient as

$$D_p = \theta\left(\frac{c}{\tau^2}\right) D \qquad (7)$$

For studies of tuffs from the Neveda test site Walter (1982)

found the term $c/\tau$ ranged from about 0.1 to 0.3, although he did attempt to separately evaluate each parameter.

For practical purposes in the analysis of matrix diffusion it is sufficient to consider $c/\tau$ as an integral part of the effective diffusion coefficient, $D^*$, in the porous media. A laboratory or field diffusion experiment will most likely yield information which can most readily provide $D_p$. However for fracture transport and matrix diffusion calculations the porosity should remain a spatial variable (Grisak and Pickens, 1980) to allow for the differences in porosity between a fracture (where under ideal parallel plate conditions $\theta = 1.0$) and the rock matrix. We can therefore define the effective diffusion coefficient in the matrix, $D^*$, as

$$D^* = \frac{c}{\tau^2} D \tag{8}$$

and the resulting one-dimensional (y direction) porous rock matrix diffusion equation (modification of Fick's second law of diffusion) is

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial y}\left(\theta D^* \frac{\partial C}{\partial y}\right) \tag{9}$$

APPENDIX 3

Guidelines
for
in-situ pressure measurement
with Lynes Hydrologic Tool and
Single Packer Configuration
(Reproduced from GTC, 1983)

## INTRODUCTION

These guidelines have been prepared specifically for the NAGRA drilling program and are directly related to the Lynes hydrologic tools and single packer configurations currently in use on site. The single packer tests are conducted for the most part during drilling in a recently drilled 5 to 15 meter interval.

The single packer test is conducted during drilling because there are limits to the ability to measure representative pressures after the borehole has been open and intervals have been subjected to annulus pressure for extended lengths of time. In a single packer test, the drill string is removed from the borehole and the packer is lowered on tubing to the selected depth. The packer is inflated and the test interval is thereby isolated between the packer and the bottom of the borehole.

### Objectives of Single Packer Tests

- The primary objective of a single packer test is to <u>measure the in-situ pressure in the test interval</u>.

Other information is also obtained during a test, such as estimates of hydraulic parameters and information regarding the potential of the interval for geochemical sampling; however, these are complementary and secondary to the primary objective of pressure measurement.

Although the guidelines have been prepared in the context of the NAGRA program, they are sufficiently general in nature that, providing the initial condition of an instantaneous pressure increase above both the annulus (overflow) pressure and the estimated static pressure can be achieved, the guidelines can be employed in virtually any test program. The procedures chart and schematic response curves form the basis of the guidelines. The chart and curves are cross-referenced using the times ($t_1$, $t_2$ etc.)

and each box in the chart is subscripted by the possible response curves which could have led to that point in the chart. The times ($t_1$, $t_2$, $t_3$, $t_4$, $t_{max}$) are directly related to:

- the maximum allowed testing period;
- the magnitude and shape of the equipment compliance function;
- the temperature differential between the drilling fluid and the rock and the length of time the interval has been subjected to drill fluid temperature;
- the pressure differential between the drilling annulus overflow and the in-situ pressure and the length of time the interval has been subjected to annulus pressure.

With the above conditions defined, a minimum permeability can be derived which will allow correct pressure measurement, and the times $t_1$, $t_2$, $t_3$, and $t_4$ can be assigned. However, some estimation and technical judgement is required. In most instances, there are some relatively permeable zones (i.e. the granite/ Buntsandstein contact and the upper weathered portion of the granite) which can provide accurate and reproducible pressure measurements. These pressure measurements provide the basis for estimating the in-situ pressure ($P_{static\ est}$) in subsequent intervals. Other observations are also employed to estimate the pressure, such as drill fluid monitoring (i.e. changes in rate of inflow or loss and drill water tracer data). However, even if the in-situ pressure is exceptionally high or low compared to the estimated pressure, there is allowance in the procedures to ensure that these pressures are obtained and that they are not temperature or compliance related.

The times ($t_1$, $t_2$, $t_3$, $t_4$) recommended for the Nagra crystalline rock drilling program are based on analysis and interpretation of field data and conditions observed to date, in conjunction with numerous theoretical calculations over the range of permeabilities, pressures, and temperatures likely to be encountered. These analyses and calculations are presented in the GTC (1983) report to Nagra "Pressure Measurement at Depth". The report addresses fluid properties and pressure/temperature relationships as well as equipment compliance and related effects. Also presented are calculations related to borehole history and length of time an interval has been open to flowing annulus or drilling fluid overflow pressure.

The recommended times, based on the above analyses and calculations and a maximum test period (starting at $t_0$) of 20 hours are:

- $t_1$ 1 hour
- $t_2$ 2 hours
- $t_3$ 6 hours
- $t_4$ 16 hours
- $t_{max}$ 20 hours

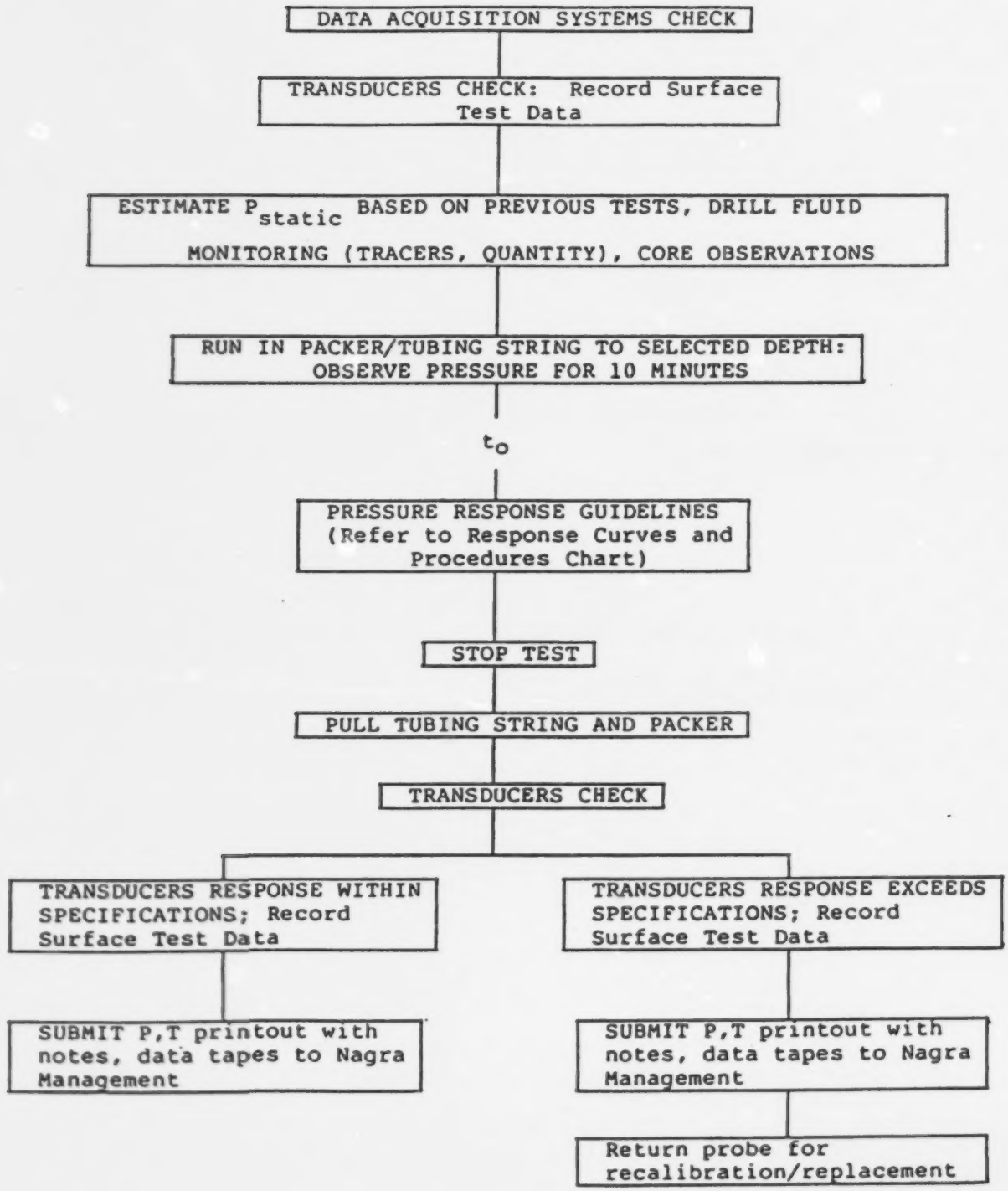
A confirmatory pulse or slug injection or withdrawal test is performed at the completion of every pressure measurement. The injection or withdrawal is designed to approach the measured in-situ pressure from the opposite direction of the shut-in pulse, thereby reducing uncertainty in the pressure measurement. The withdrawal or injection tests should be completed within the 20 hour period ($t_{max}$). Continuation of the test beyond $t_{max}$ would be done on direction of NAGRA management. Otherwise all test procedures should be completed with times $t_2 < t < t_{max}$, or between 2 and 20 hours.

A flow chart for the general test procedures follows. There is no allowance for malfunction or down time. Should equipment

malfunction occur at any point above STOP TEST in the flow chart, an estimate of the time required to correct the problem should be made and management consulted for a decision on whether to resume drilling or effect repairs.

The pressure response guidelines (Procedures Chart (Figure 1) and Response Curves (Figures 2 and 3)) are provided in the pocket immediately following the flow chart.

**IN-SITU PRESSURE MEASUREMENT WITH LYNES HYDROLOGIC TOOL AND SINGLE PACKER CONFIGURATION**


DATA ACQUISITION SYSTEMS CHECK
TRANSDUCERS CHECK: Record Surface Test Data
ESTIMATE $P_{static}$ BASED ON PREVIOUS TESTS, DRILL FLUID MONITORING (TRACERS, QUANTITY), CORE OBSERVATIONS
RUN IN PACKER/TUBING STRING TO SELECTED DEPTH: OBSERVE PRESSURE FOR 10 MINUTES
$t_o$
PRESSURE RESPONSE GUIDELINES (Refer to Response Curves and Procedures Chart)
STOP TEST
PULL TUBING STRING AND PACKER
TRANSDUCERS CHECK
TRANSDUCERS RESPONSE WITHIN SPECIFICATIONS; Record Surface Test Data
TRANSDUCERS RESPONSE EXCEEDS SPECIFICATIONS; Record Surface Test Data
SUBMIT P,T printout with notes, data tapes to Nagra Management
SUBMIT P,T printout with notes, data tapes to Nagra Management
Return probe for recalibration/replacement

PROCEDURES CHART

The procedures chart for in-situ pressure measurement addresses the interrelationships between pressure, temperature and equipment compliance as they relate to the possible observed pressure responses following interval isolation and an initial "shut-in squeeze pressure". The procedures are somewhat different depending on whether the estimated static pressure in the formation is above or below the annulus pressure (Figures 2 and 3 respectively). The annulus pressure refers to the downhole pressure corresponding to the elevation of the overflow of drilling fluid at the wellhead.

The procedures chart is self-explanatory except for definition of the terms and abbreviations. The following describes the terminology and defines the abbreviations utilized in the procedures chart.

- $t_{1,2,3,4}$ times corresponding to the schematic response curves.
- $P_{static\ est}$ the estimated static pressure (absolute) in the interval to be tested; based on previous tests, drill fluid monitoring (quantity, tracers), core observations.
- $P_{annulus}$ the downhole pressure (absolute) corresponding to the elevation of the overflow of drilling fluid at the wellhead.
- $P_{SI}$ the pressure (absolute) resulting from the "shut-in squeeze".
- $\Delta P$ the change in pressure at the proceeding time designation, above the illustrated box. For example, procedures and $\Delta P$'s between $t_2$ and $t_3$ on the procedures chart relate to the time

$t_2$. $\Delta P$ does not refer to the change in pressure between $t_2$ and $t_3$ but rather refers to the slope and magnitude of the pressure change at $t_2$.

$\Delta P > 0$, indicates an increase in pressure
$\Delta P < 0$, indicates a decrease in pressure
$\Delta P \approx 0$, indicates approximately 0

- $\Delta T$ — the change in temperature at the proceeding time designation, above the illustrated box. See $\Delta P$ for complete explanation.

- $K_{req}$ — the permeability required to obtain 85% recovery from the shut-in squeeze within the allowed testing time. For the NAGRA testing program the following have been derived

| Borehole Open | $K_{req}$ |
|---|---|
| 1 day | $10^{-11}$ m/s |
| 2 days | $10^{-10}$ m/s |

- $K_{min}$ — the permeability calculated from the early time response data. Includes equipment compliance and temperature effects.

- PULSE/SLUG WITHDRAWAL/ INJECTION TEST — an instantaneous decrease or increase in the pressure or water level in the test interval, with subsequent monitoring of pressure response or water level recovery.

PRESSURE RESPONSE CURVES

Two schematic response curve diagrams are provided; one with $P_{static\ est}$ above $P_{annulus}$ and one with $P_{static\ est}$

below $P_{annulus}$. The differences are concerned with system checks related to packer bypass or shut-in valve leaks. The test procedures vary only at $t_3$.

For discussion purposes, the response curves of $P_{static\ est} > P_{annulus}$ will be used. There are 5 "master curves" illustrated which can possibly develop in response to the shut-in squeeze pressure. The squeeze pressure is important because it provides a common point of departure and, since it is always above $P_{static\ est}$ and $P_{annulus}$, it serves to limit the possible responses. The master curves and their subsets are briefly explained below.

Master Curve 1

Segment B

| | |
|---|---|
| 1 | could be responding to high in-situ pressure in high or low K interval; pressure increase could be due to equipment compliance or rising temperature in a low K interval |

Segment C

| | |
|---|---|
| 1 | responding to high in-situ pressure |
| 1a | new in-situ pressure |
| 1b | compliance/temperature response has peaked, however $K_{min}$ is not sufficiently high to achieve $P_{static\ est}$ in the allowed time |
| 1c | compliance temperature response has peaked and $K_{min}$ is high enough to continue the test |

Segment D

| | |
|---|---|
| 1 | responding to high in-situ pressure |
| 1.1, 1c.1 1c.3, 1c.5 | new in-situ pressure |
| 1c.2 | measured pressure = $P_{static\ est}$ |

| | |
|---|---|
| 1c.4 | measured pressure = $P_{annulus}$; requires system check |
| 1c.6 | in-situ pressure appears very low; requires more time for pressure measurement |

Segment E

| | |
|---|---|
| 1, 1c.6 | new in-situ pressure |

Master Curve 2

Segment B

| | |
|---|---|
| 2 | could be correct in-situ pressure, or pressure remains elevated due to compliance/temperature effects, or permeability is very low |

Segment C

| | |
|---|---|
| 2 | new in-situ pressure is greater than $P_{SI}$ or temperature is increasing |
| 2a | new in-situ pressure |
| 2b | compliance/temperature effects have peaked, or were not present, however $K_{min}$ is not sufficiently high to continue |
| 2c | compliance/temperature effects have peaked, or were not present, and $K_{min}$ is high enough to continue |

Segment D

| | |
|---|---|
| 2 | responding to high in-situ pressure |
| 2.1, 2c.1 2c.3, 2c.5 | new in-situ pressure |
| 2c.2 | measured pressure = $P_{static\ est}$ |
| 2c.4 | measured pressure = $P_{annulus}$; requires system check |
| 2c.6 | in-situ pressure appears very low; requires more time for pressure measurement |

Segment E

| | |
|---|---|
| 2, 2c.6 | new in-situ pressure |

Master Curve 3

Segment B

| | |
|---|---|
| 3 | presssure dropping slowly, could be low K and/or compliance/temperature effects |

Segment C

| | |
|---|---|
| 3 | compliance effects are minimal. Pressure could be rising due to temperature, or possible borehole "skin" effect (pressure around borehole during drilling is lower than in-situ pressure, therefore a low pressure "skin" has developed) |
| 3a | new in-situ pressure |
| 3b | possible continued temperature effects, however $K_{min}$ is not sufficiently high to continue |
| 3c | possible continued temperature effects, however $K_{min}$ is high enough to continue |

Segment D

| | |
|---|---|
| 3 | responding to high in-situ pressure |
| 3.1, 3c.1 3c.3, 3c.5 | new in-situ pressure |
| 3c.2 | measured pressure = $P_{static\ est}$ |
| 3c.4 | measured pressure = $P_{annulus}$; requires systems check |
| 3c.6 | in-situ pressure appears very low; requires more time for pressure measurement |

Segment E

| | |
|---|---|
| 3, 3c.6 | new in-situ pressure |

Master Curve 4

Segment B

| | |
|---|---|
| 4 | pressure dropping relatively rapidly; could be either intermediate K or relatively high K with compliance and temperature effects |

Segment C

| | |
|---|---|
| 4 | compliance effects are minimal. Pressure could be rising due to temperature or possible borehole "skin" effect (pressure around borehole during drilling is lower than in-situ pressure, therefore a low pressure "skin" has developed) |
| 4a | new in-situ pressure |
| 4b | possible temperature or "skin" effects reduce rate of pressure drop; $K_{min}$ is not sufficiently high to continue |
| 4c | possible continued temperature effects, however $K_{min}$ is high enough to continue |

Segment D

| | |
|---|---|
| 4 | responding to high in-situ pressure |
| 4.1, 4c.1 4c.3, 4c.5 | new in-situ pressure |
| 4c.2 | measured pressure = $P_{static\ est}$ |
| 4c.4 | measured pressure = $P_{annulus}$; requires system check |
| 4c.6 | in-situ pressure appears very low; requires more time for pressure measurement |

Segment E

| | |
|---|---|
| 4, 4c.6 | new in-situ pressure |

Master Curve 5

Segment B

| | |
|---|---|
| 5 | indicates very high K, compliance or temperature effects are minimal |

Segment C

| | |
|---|---|
| 5, 5b, 5d | new in-situ pressure |
| 5a | measured pressure = $P_{static\ est}$ |
| 5c | measured pressure = $P_{annulus}$; requires system check |
| 5e | in-situ pressure appears very low; requires more time for pressure measurement |

Segment D

| | |
|---|---|
| 5e.1 | new in-situ pressure |
| 5e.2 | in-situ pressure appears very low; requires more time for pressure measurement |

Segment E

| | |
|---|---|
| 5e.2 | new in-situ pressure |

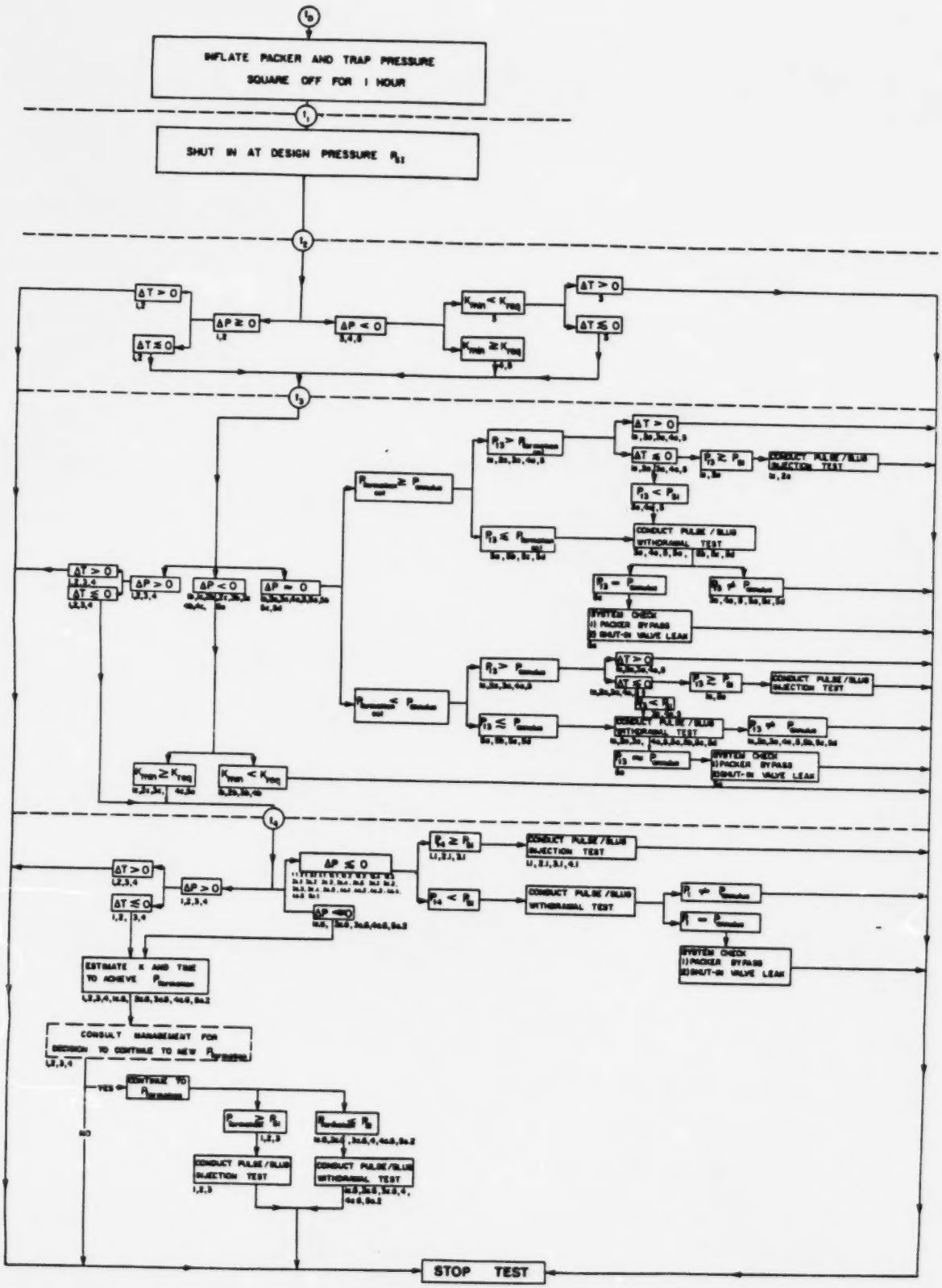
ENCLOSURES (Pocket)

Figure 1. Procedures for In-Situ Pressure Measurement

Figure 2. Pressure Response Curves: $P_{static\ est} > P_{annulus}$

Figure 3. Pressure Response Curves: $P_{static\ est} < P_{annulus}$

Figure 1. Procedures for In-Situ Pressure Measurement



## LEGEND - For Figure 1.

- $t_{1,2,3,4}$ — times corresponding to the schematic response curves.

- $P_{formation\ est}$ — the estimated formation pressure in the interval (prior to the drilling/testing disturbances) based on previous tests, drill fluid monitoring (outflow/inflow quantity, tracer concentration) and core observations.

- $P_{formation}$ — new estimate of formation pressure determined from the test.

- $P_{annulus}$ — the downhole pressure corresponding to the elevation of the overflow of drilling fluid at the wellhead.

- $P_{SI}$ — the pressure in the test interval at time of shut-in, resulting from all influences, including pressure prior to shut-in, "shut-in squeeze", applied load, compliance, etc. This pressure is chosen before the start of the test based on the annulus pressure, an estimate of the anticipated formation pressure and the "designed" pressure pulse differential required for the pulse test. This pressure is also the starting pressure for any subsequent test analysis.

- $\Delta P$ — the change in pressure at the preceeding time designation, above the illustrated box. For example, procedures and $\Delta P$'s between $t_2$ and $t_3$ on the procedures chart relate to the time $t_2$. $\Delta P$ does <u>not</u> refer to the change in pressure between $t_2$ and $t_3$ but rather refers to the slope and magnitude of the pressure change <u>at</u> $t_2$.

  $\Delta P > 0$, indicates an increase in pressure

  $\Delta P < 0$, indicates a decrease in pressure

  $\Delta P = 0$, indicates approximately no change in pressure.

- $\Delta T$ — the change in temperature at the preceeding time designation, above the illustrated box. For example, procedures and $\Delta T$'s between $t_2$ and $t_3$ on the procedures chart relate to the time $t_2$. $\Delta T$ does not refer to the change in temperature between $t_2$ and $t_3$ but rather refers to the slope and magnitude of the pressure change <u>at</u> $t_2$.

  $\Delta T > 0$, indicates an increase in temperature

  $\Delta T < 0$, indicates a decrease in temperature

  $\Delta T = 0$, indicates approximately no change in temperature

- $K_{req}$ — the hydraulic conductivity required to obtain 85% recovery from $P_{SI}$ within $t_{max}$. $K_{req}$ is approximated using the estimated $P_{formation\ est}$ and the length of time the interval has been open to annulus pressure (see section 3.8.1 for details). For a $t_{max}$ of 20 hours and a borehole open period up to 48 hours, $K_{req}$ is about $10^{-10}$ m/s.

- $K_{min}$ — the hydraulic conductivity required to obtain 85% recovery from $P_{SI}$ within $t_{max}$. $K_{min}$ is used for direct comparison to $K_{req}$. $K_{min}$ is estimated from the pressure response data obtained between $t_1$ and $t_2$ and between $t_2$ and $t_3$, under the conditions where $\Delta P < 0$. The estimate is made using standard type curve (hydraulic response) analysis on the pressure data. If temperature effects are significant with $\Delta T > 0$, $K_{min}$ will be less than $K_{req}$, resulting in "STOP TEST".

- PULSE/SLUG WITHDRAWAL/ INJECTION TEST — an instantaneous decrease or increase in the pressure or water level in the test interval, with subsequent monitoring of pressure response of water level recovery monitoring of pressure response or water level (see sections 2.3 and 2.4).

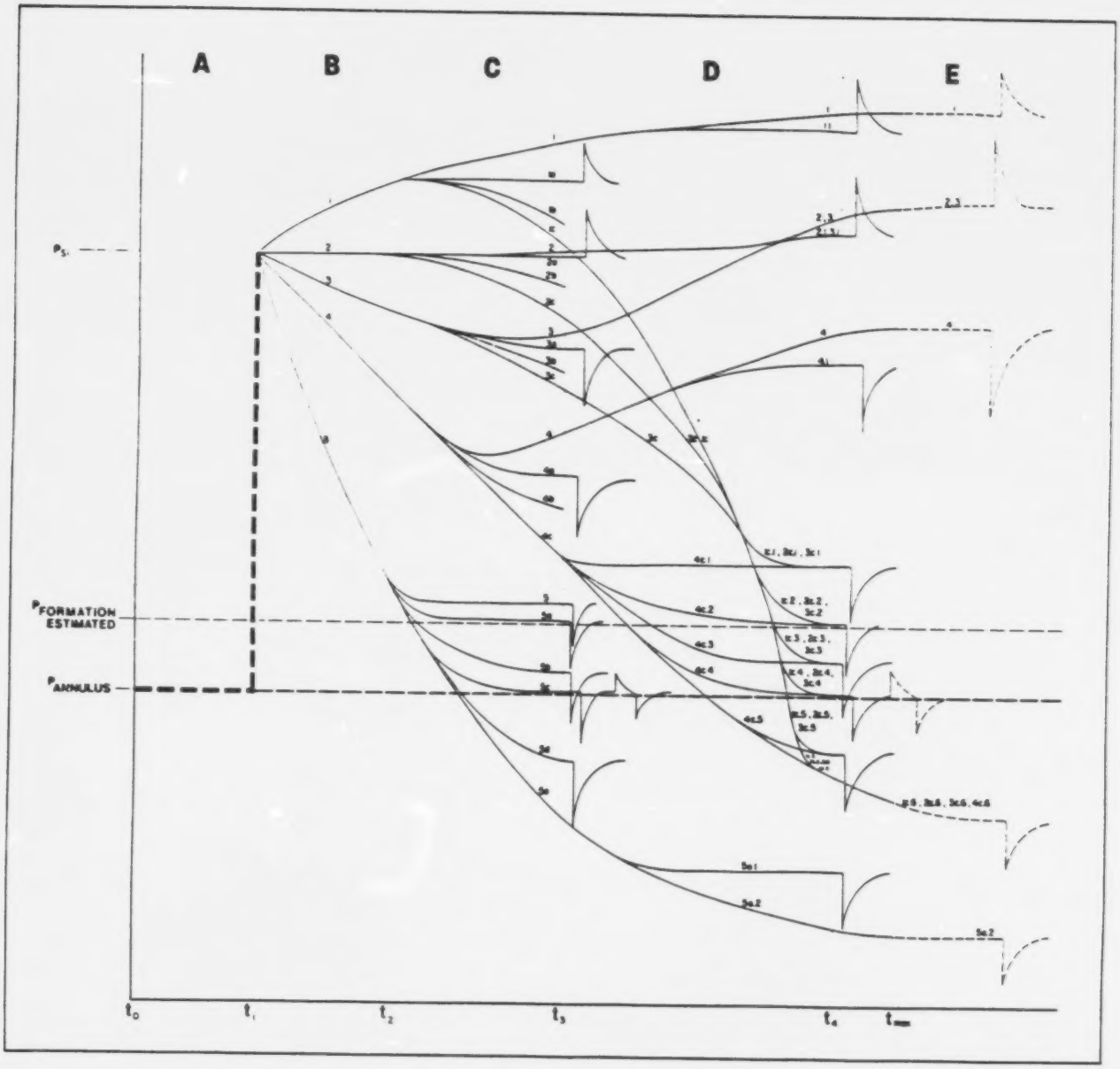
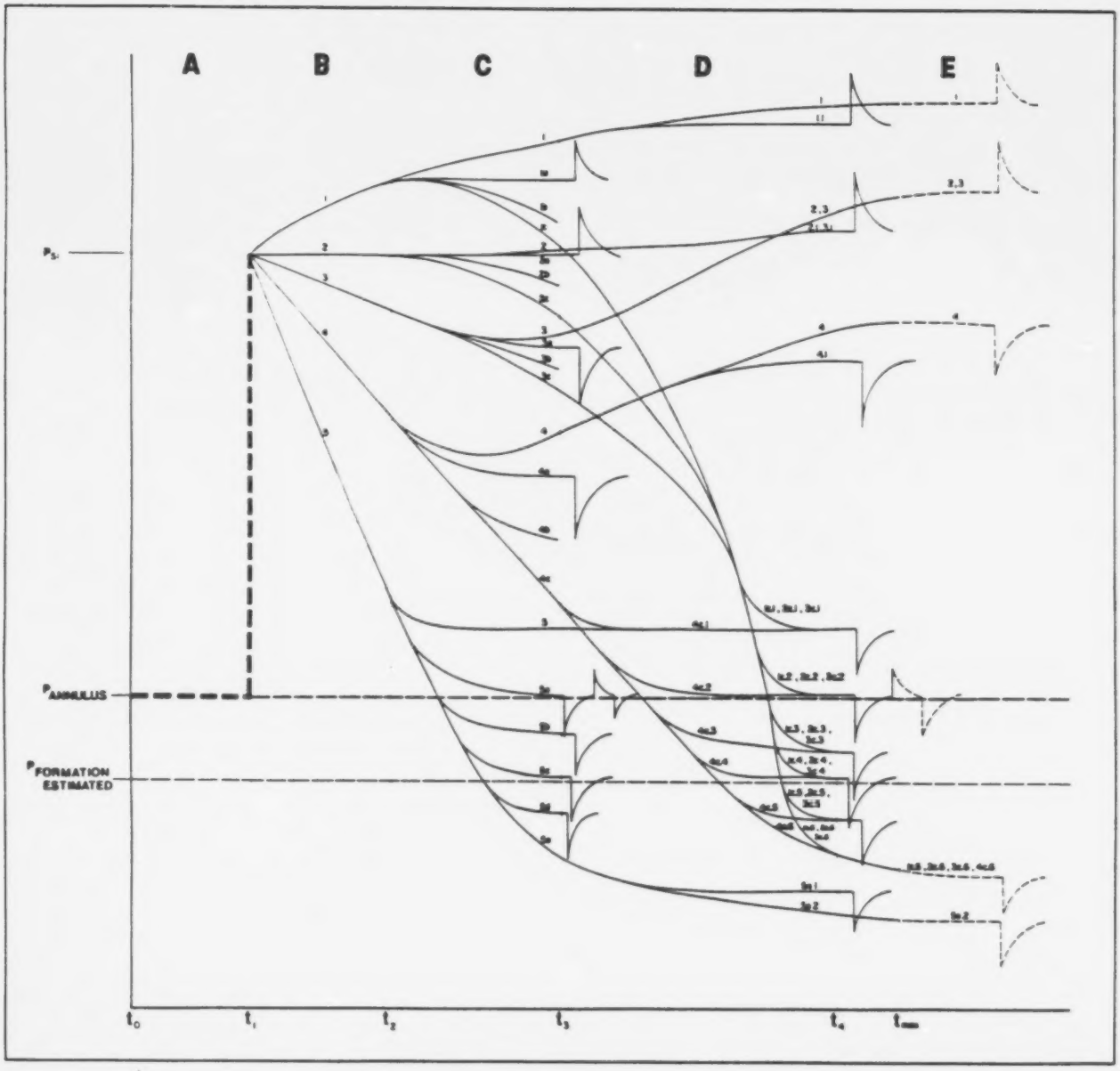



Figure 2. Pressure Response Curves: $P_{static\ est} > P_{annulus}$




PRESSURE RESPONSE CURVES $P_{FORMATION\ EST} < P_{ANNULUS}$

Figure 3. Pressure Response Curves: $P_{static\ est} < P_{annulus}$

APPENDIX 4

# IAEA SEMINAR ON THE SITE INVESTIGATION TECHNIQUES AND ASSESSMENT METHODS FOR UNDERGROUND DISPOSAL OF RADIOACTIVE WASTES

Trip Report to
Atomic Energy Control Board

Prepared by

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February 29, 1984

## Summary

A seminar entitled "Site Investigation Techniques and Assessment Methods for Underground Disposal of Radioactive Wastes" was sponsored by the International Atomic Energy Agency in Sophia, Bulgaria during Februry 6 - 10, 1984. The seminar was organized into the following sessions:

(1) Overview and general approaches
(2) Regional and local site investigations
(3) Detailed geoscientific investigations
(4) Subsurface techniques for site investigations
(5) Hydrogeological studies
(6) Studies on geochemical and physiochemical parameters
(7) Investigation of properties of ground water
(8) Site assessment and confirmation studies

Several of the national programs involved in geologic research for the disposal of radioactive wastes are advanced with respect to the development and utilization of comprehensive site investigation/ assessment methodologies. Most notable among these are Canada, Finland, France, South Africa, Sweden, Switzerland and the United States. The applicability of many of the investigation and assessment techniques are site-specific and therefore the techniques chosen for a particular site must be evaluated and screened prior to and during the site investigation. Correlation of information obtained from various methods can allow an optimization of the continuing and future site investigations. The various techniques for assessing the hydrogeologic characteristics of a site are summarized in the attached tables:

Table 1. Borehole Investigation Methods
Table 2. Surficial Investigation Methods
Table 3. Remote Sensing Techniques

Table 4. Laboratory Methods for Testing Borehole Cores
Table 5. Summary of Physical, Chemical, Isotopic and Bacteriological Parameters from Ground Water Samples

Table 1. Borehole Investigation Methods

| Borehole Investigation Method | Information Obtained |
|---|---|
| Drill Core | Rock types<br>Fracture orientation<br>Geological/mineralogical data<br>Fracture infilling materials |
| Drill fluid logging | Formation gases<br>Fluid gains/losses at permeable zones (e.g., fracture zones) |
| Hydraulic tests<br>- drill stem tests<br>- pulse tests<br>- slug tests<br>- constant head tests<br>- injection or pumping tests (constant or variable rate)<br>- cross-hole interference | Hydraulic conductivity<br>Identification of fracture zones, permeable zones and lower permeability zones<br>In-situ formation pressure (in some cases) |
| Hydrogeologic reconnaissance log (hyraulic tests) | Hydraulic conductivity profile of borehole |
| Labelled slug test | Flow rates in borehole<br>Inflow/outflow zones |
| Fluid logging | Inflow/outflow zones |
| Spinner flowmeter test | Flow rates in borehole<br>Inflow/outflow zones |
| Tracer tests | Fracture interconnectivity<br>Porosity estimate |
| Self (Spontaneous) log | Formation water properties<br>Lithology |
| High resolution dipmeter | Formation dip and direction<br>Stratigraphic interpretation<br>Hole geometry |
| Electromagnetic propagation log | Porosity variation |

Table 1. (contiued)

| Borehole Investigation Method | Information Obtained |
|---|---|
| Resistivity logs | Lithology<br>Drilling fluid invasion |
| Cross-hole electrical | Fracture zones |
| Density | Porosity; lithology |
| Neutron | Porosity; lithology |
| Sonic | Fracture location |
| Long-spaced sonic | Porosity; lithology<br>Fracture location |
| Natural gamma ray | Lithology |
| High resolution temperature log | Geothermal gradient<br>Inflow/outflow zones |
| Continuous directional survey | Azimuth and deviation from vertical of borehole |
| Sidewall sampler | Samples from wall of borehole |
| Caliper log | Hole size information |
| Borehole geometry tool | Hole size information |
| Sonic televiewer log | Fracture orientation |
| Tube wave | Fracture location |
| TV log | Fracture frequency and indication of whether open or closed |
| Downhole geochemical probe | Eh<br>pH<br>Temperature |
| Multipacker pressure measurement system | Formation pressures at various depths |

Table 2. Surficial Investigation Methods

| Method | Information Obtained |
|---|---|
| Seismic survey | Formation depths and thicknesses<br>Depth to crystalline rock surface |
| Gravity survey | Structural<br>Depth to crystalline rock surface |
| Sampling of spring discharges | Structural |
| Electromagnetic | Structural |
| Resistivity | Formation depths and thicknesses |
| Mapping outcrops | Structural<br>Fracture zone data |

Table 3. Remote Sensing Techniques

| Method | Information Obtained |
|---|---|
| Aeromagnetic survey | Variations in surficial geology<br>Magnetic anomalies associated with faults |
| Airborne electromagnetic | Variations in surficial geology<br>Conductors associated with ground water discharge at faults |
| Thermal infrared line scanning | Variations in surficial geology<br>Soil moisture variation<br>Structural information in areas with little overburden materials<br>Ground water discharge zones |
| Landsat imagery | Structural information on large scale<br>Faults |

Table 4. Laboratory Methods for Testing Borehole Cores

| Method | Information Obtained |
|---|---|
| Mineralogical analysis | Rock type<br>Minerals in rock matrix<br>Fracture minerals |
| Fracture analysis | Fracture orientation and frequency |
| Diffusion experiments | Solute diffusion coefficients in rock matrix |
| Porisimetry techniques | Porosity |
| Permeability | Permeability |
| Sorption experiments | Retardation capabilities of rock matrix |
| Thermal experiments | Thermal conductivity<br>Specific heat; heat capacity<br>Thermal expansivity |

Table 5. Summary of Physical, Chemical, Isotopic and Bacteriological Parameters from Ground Water Samples

| Parameter | Item |
|---|---|
| Physical/chemical parameters | Temperature<br>Electrical conductivity<br>pH<br>Redox potential<br>Natural radioactivity<br>Total organic carbon |
| Chemical analyses | $Na^+$, $K^+$, $Mg^{2+}$, $Ca^{2+}$, $Cl^-$, $HCO_3^-$, $CO_3^{2-}$, $SO_4^{2-}$, $H_2SiO_3$, $NH_4^+$, $NO_3^-$, $NO_2^-$, $Mn^{2+}$, $Fe^{2+}$, $Fe_{total}$, $Li^+$, $Sr^{2+}$, $H_3BO_3$, Al , $F^-$, $Br^-$, $I^-$, $Rb^+$, $Cs^+$, $Ba^{2+}$, Pb, P, As, Cu, Ni, Se, Zn, U, Ra, sulphide, phosphate. |
| Gases | $O_2$, $CO_2$, $H_2S$, $N_2$, Ar, He, $H_2$, $CH_4$, $C_2H_6$ |
| Bacteriological | |
| Isotopic | $^{13}C$, $^{14}C$, $^{18}O$, $^{3}H$, $^{36}Cl$, $^{234}U/^{238}U$, Helium, Neon, Argon, Kryton, Xenon, $^{81}Kr$, $^{230}Th$, $^{232}Th$. |